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S. A. BILGRAMI *Pakistan Chrome Mines Limited, Hindubagh, West Pakistan*

Distribution of Cu, Ni, Co, V, and Cr in Rocks of the Hindubagh Igneous Complex, Zhob Valley, West Pakistan

Abstract: The writer has made colorimetric and spectrochemical determinations on six specimens of ultramafic and mafic rocks from the Hindubagh igneous complex for which bulk chemical analyses are available. He has concluded that the distribution of five trace elements, Cu, Ni, Co, V, and Cr, in the Hindubagh rocks is normal. The results of the two analytical methods agree closely.

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INTRODUCTION

Hindubagh is a small town about 74 miles northeast of Quetta in the former Baluchistan, Pakistan (Fig. 1). The area investigated comprises about 90 square miles (Survey of Pakistan topographical sheet nos. 34N/13 and 34N/14) and is one of very rugged topography. The highest point, Saplaitorgarh, is 8361 feet above sea level, and the average altitude of the plains is 6000 feet.

The Hindubagh igneous complex includes serpentinites, dunites, chromitites, harzburgites, and peridotites. The metamorphic rocks are crystalline limestones, hornblende schists, and gneisses, and the sedimentary rocks are shales and limestones. All are cut by dolerite dikes. Little is known of the mineralogy and petrology of the area; the writer (1961) has given a preliminary description of this igneous complex elsewhere (Bilgrami, 1961). Further work on the geology of the Zhob Valley complex is in progress; this paper is a part of the regional study.

The writer analysed six rock samples for which bulk chemical analyses are available for Cu, Ni, Co, V, and Cr by colorimetric and spectrochemical methods. The main objects were to determine if any systematic variations exist between the major and minor elements in the rocks of this complex and to compare the results of the two methods.

ACKNOWLEDGMENTS

Sincere thanks are due to Prof. E. B. Sandell for his help and advice with analytical techniques, to Dr. S. S. Goldich for his interest in the work and a critical reading of the manuscript, and to J. Ito and Harry J. Rose, Jr., for help with spectrochemical techniques. A Ful-

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bright travel grant, a Smith-Mundt Fellowship, and permission to use research and library facilities at the University of Minnesota and Harvard University are gratefully acknowledged.

and grinding were done in agate mortars using acetone as liquid medium. Further grinding was done on dried samples to drive out all traces of acetone.

For the colorimetric determinations, the

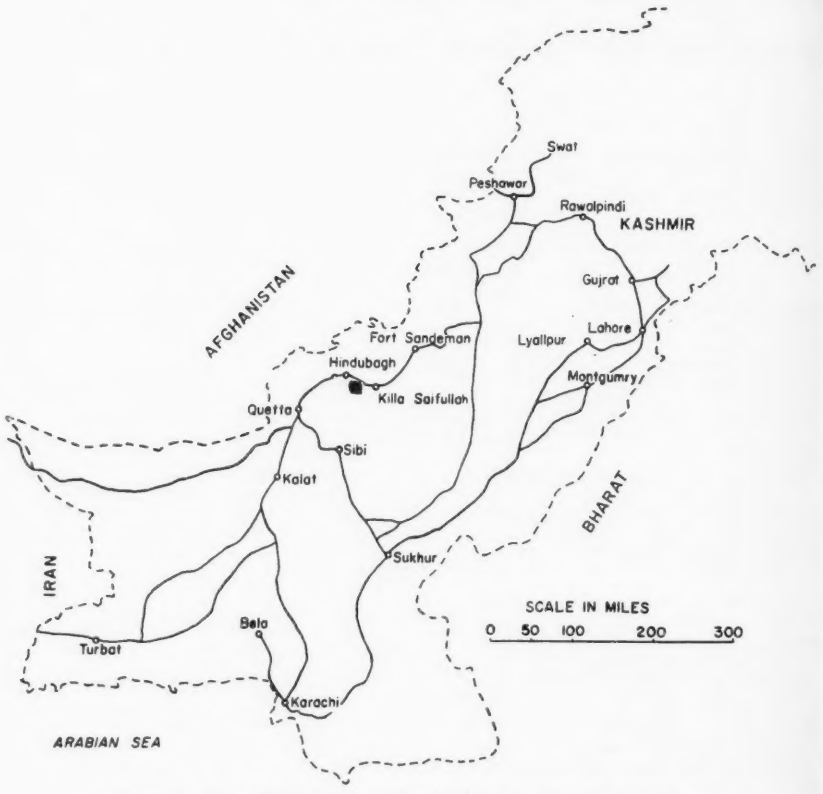

Figure 1. Map of West Pakistan showing location of the Hindubagh area

EXPERIMENTAL WORK

The writer crushed 25–30 gms of each of the rock samples in steel and agate mortars and sieved the material through silk to avoid any contamination from brass sieves. All samples were crushed to less than 120 mesh for colorimetric work. For spectrochemical work the writer diluted the samples 2–4 times with a mixture of SiO_2 (80 per cent), Na_2CO_3 (10 per cent), and graphite (10 per cent). Mixing

Table 1. Distribution of Cu, Ni, Co, V, and Cr in G–1 and W–1 (PPM)
Analyst, S. A. Bilgrami

	Cu		Ni		Co		V		Cr	
G–1	10	11	4	4	2	2	20	20	18	18
	12		3		3		20		19	
W–1	115	116	75	75	35	35	255	256	120	122
	120		72		34		248		124	
	114		78		36		256		125	
	115		76		35		262		120	

writer used the techniques Sandell (1959) described. In all cases the writer took 0.25 gm of sample and made duplicate determinations on separate portions of the samples. As a preliminary check on the working curves the writer made duplicate determinations of Cu, Ni, Co, V, and Cr on G–1 and W–1. In the case of W–1 the writer used two different samples supplied by S. S. Goldich and E. B. Sandell (masked sample) and hence made four determinations of each of the above elements listed on W–1. Table 1 shows the results; they agree well with those that Young (1958) presented.

The writer made the spectrochemical determinations on a Baird Associates 2-m grating spectrograph using 220-volt d.c. arc at 12 amperes current. A rotating sector was placed in front of the slit to effectively reduce the exposure to 30 per cent of the total. The slit was set at 25 microns. The electrode gap was maintained manually at 4 mm. Twenty mg of the standard or sample was transferred to the sample-carrying electrode, which became the anode during excitation. The arcing was started at 3 amperes, and after 10 seconds the current was raised to 12 amperes; all samples were arced to completion; the arcing time was 2½ minutes. The developed plates gave a transparency of 85–90 per cent on an Associated Research Laboratories nonrecording densitometer, on which all plates were read.

The writer prepared two bulk reference samples (1A and 2A) by averaging analyses 1 and 2 (Table 2) and analyses 3–6 (Table 2) and recalculating these values to 100 per cent on a water-free basis (Table 3). Known quantities of Cu, Ni, Co, V, and Cr were added to the samples, and replicate arcing was done for making standard plates. In addition, G–1 and

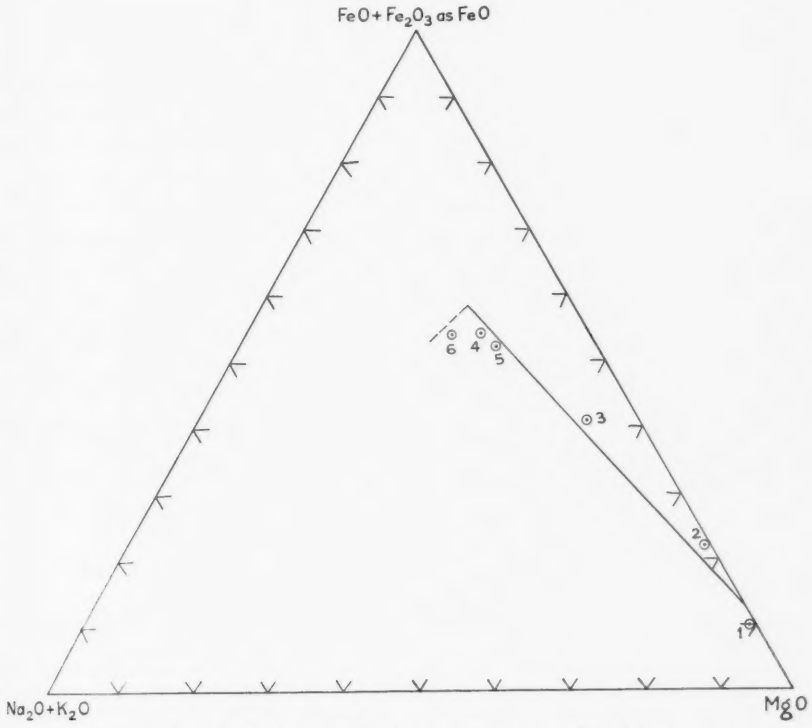

Figure 2. Trend of variation in the chemical composition of Hindubagh rocks

TABLE 2. CHEMICAL AND MINERALOGICAL COMPOSITIONS OF HINDUBAGH ROCKS
Analyst, S. A. Bilgrami

	(1)	(2)	(3)	(4)	(5)	(6)
SiO_2	37.75	41.22	40.05	49.61	48.81	50.33
Al_2O_3	2.17	4.47	16.12	15.55	15.71	15.83
TiO_2	0.04	0.41	0.84	0.86	0.82	0.91
Fe_2O_3	2.94	4.58	4.42	6.54	1.70	8.57
FeO	2.37	5.62	8.21	4.72	9.94	2.62
MnO	0.16	0.05	0.06	0.03	0.17	0.02
MgO	43.42	34.69	15.52	5.86	7.54	5.57
CaO	trace	2.84	7.69	10.56	9.32	10.01
Na_2O	0.31	0.32	1.28	2.33	1.97	2.51
K_2O	0.12	trace	0.84	0.52	0.95	1.01
H_2O+	9.76	5.47	4.43	3.11	3.18	2.14
H_2O-	0.46	0.32	0.08	0.05	0.07	0.05
P_2O_5	0.12	0.11	0.42	0.18	0.12	0.22
	99.62	100.10	99.96	99.92	100.30	99.80

(1) Dunite (specimen 308). Fine-grained, olive green, composed of olivine (51.19 per cent), serpentine (46.77 per cent), spinel (2.03 per cent)

(2) Peridotite (specimen 2). Coarse-grained, dark, composed of olivine (34.54 per cent), augite (12.05 per cent), enstatite (26.04 per cent), serpentine (22.04 per cent), spinel (5.51 per cent)

(3) Hornblende peridotite (specimen 318). Coarse-grained, dark, composed of olivine (15.88 per cent), augite (16.04 per cent), brown hornblende (22.20 per cent), biotite (6.85 per cent), plagioclase (3.10 per cent), chlorite (16.06 per cent), serpentine (18.53 per cent), spinel (1.06 per cent)

(4.) Normal dolerite (specimen 245). Medium-grained, dark, composed of plagioclase (49.93 per cent), augite (35.40 per cent), chlorite (11.13 per cent), spinel (3.43 per cent)

(5) Green hornblende dolerite (specimen 312). Medium-grained, dark, composed of augite (21.98 per cent), green hornblende (55.93 per cent), sericite (21.99 per cent)

(6) Quartz dolerite (specimen 287). Medium-grained, light-colored, speckled rock composed of augite (14.10 per cent), green hornblende (4.56 per cent), sericite (54.24 per cent), chlorite (22.75 per cent)

TABLE 3. CHEMICAL COMPOSITIONS OF REFERENCE SAMPLES

	G–1	W–1	(1)	(1A)	(2)	(2A)
SiO_2	72.79	52.66	39.48	42.97	47.25	48.88
Al_2O_3	14.26	15.01	3.32	3.68	15.80	16.44
TiO_2	0.26	1.08	0.27	0.29	0.78	0.80
Fe_2O_3	0.78	1.41	3.76	4.09	5.30	5.49
FeO	0.97	8.73	3.99	4.43	6.37	6.59
MnO	0.03	0.16	0.10	0.10	0.07	0.07
MgO	0.37	6.60	39.15	42.34	8.62	8.81
CaO	1.36	11.00	1.42	1.54	9.39	9.74
Na_2O	3.30	2.14	0.31	0.33	2.02	2.08
K_2O	5.54	0.62	0.06	0.06	0.83	0.85
H_2O+	0.25	0.45	8.00	0.00	3.29	0.00
P_2O_5	0.08	0.12	0.11	0.12	0.23	0.23
	99.99	99.98	99.97	99.95	99.95	99.98

G–1, W–1: Averages of Rock Analysis Laboratory. Analysts, Goldich and Oslund (1956, p. 814, Table 3), Westerly and Centerville

(1) Average of analyses 1 and 2, Table 2

(1A) Same as (1), recalculated to 100 per cent on water-free basis

(2) Average of analyses 3–6, Table 2

(2A) Same as (2), recalculated to 100 per cent on water-free basis

W-1 and their mixtures were also used as reference samples. The average values of trace elements for these samples (Table 1) were taken as standard values. Photographic emulsions were calibrated by the step sector method (Recommendations of ASTM Committee E-2, 1957, P. 17–20). Intensity-concentration curves were drawn on log-log graph paper from which concentration of these elements in the unknown samples were determined. The straight-line working curves produced gave no indication of self reversal or residual contamination. No background correction was necessary.

TREND OF DIFFERENTIATION

The analyses in Figure 2 and Table 2 show the range of composition exhibited by Hindubagh rocks. Fenner (1929, p. 242) has shown that in the crystallization of basaltic magmas "a final liquid rich in certain constituents, particularly iron and alkalies, and poor in silica, magnesia and lime" is produced. Walker (1930), Wager and Deer (1939), and many others have confirmed this observation. Figure 2 shows clearly that the Hindubagh rocks show a similar trend, although in addition to enrichment in iron and alkalies there is also a concomitant enrichment in silica. This trend is similar to that of the Skaergaard intrusion (although the Hindubagh rocks are considerably more mafic) and may be taken to represent something similar to what might be expected in the "Hidden Layered Series" of the Skaergaard intrusion.

The Hindubagh complex shows another notable difference compared to the Skaergaard intrusion in the absence of ferrogabbros and granophyres; however the Hindubagh area represents only a very small portion of the main Zhob valley complex (less than 1/8), and rocks similar to ferrogabbros and granophyres may be found. The writer (1961) discussed the petrogenesis of the Hindubagh complex elsewhere.

DISTRIBUTION OF TRACE ELEMENTS

In spite of the extensive researches of Goldschmidt (1934; 1937; 1938), Vogt (1921; 1923), Sandell and Goldich (1943), Nockolds and Mitchell (1948), Wager and Mitchell (1951), Shaw (1953), Ringwood (1955), Nockolds and Allen (1953; 1954; 1956), and others, knowledge of the distribution of trace elements is far from complete. Almost invariably investigators have found that the distribution of all elements in a given suite of rocks could not have been predicted, although the general trends could have been. This of course does not apply equally to all elements, and with our present knowledge the distribution of some elements can be predicted more precisely than that of others.

The distribution of trace elements in the rocks of the Hindubagh igneous complex is not unusual, and for the most part, the elements follow trends that are in keeping with present ideas of distribution.

COPPER: The concentration of copper in mafic igneous rocks is several times that in felsic igneous rocks Goldschmidt (1954, p. 178–179). It has been suggested that Cu^{+2} occurs mainly as sulfides, and various investigators have correlated the copper and sulfur contents of the rocks from various localities. Sandell and Goldich (1943, p. 175), however, have pointed out that ". . . the chemical analyses show about five times as much copper as the microscope." Grout (1910, p. 475) and Reed (1936, p. 268) have suggested that some of the copper in igneous rocks may be present in the ferromagnesian minerals. Wager and Mitchell (1951, p. 190) have postulated that in the early stages of crystallization of a magma Cu^{+1} (ionic radius 0.96 Å) and Cu^{+2} (0.72 Å) occur free and may be incorporated in silicate minerals. In these forms copper may be camouflaged by Na (0.88 Å) and Fe^{+2} (0.74 Å). In the later stages of strong fractional crystallization of the Skaergaard magma ". . . an immiscible sulfide separated out into which copper entered." The fact that Wager and Mitchell (1951) have not found enough sulfur in Skaergaard rocks to explain all the copper as sulfide phase is in conformity with the earlier observation of Sandell and Goldish (1943) that copper does enter silicate structures. In some silicate minerals of the Skaergaard intrusion copper is present. Ramdohr (1940, p. 36) is also of the opinion that some copper in the mafic rocks is present in silicates.

The writer has not observed any sulfide minerals in the Hindubagh rocks under study, and most of the copper appears to be in the silicate minerals. The rocks that formed earliest show greater concentrations of copper, although the over-all range (208–92 ppm) is not very large and is in agreement with the average figure for mafic rocks given by Sandell and Goldish (1943). Figure 3 shows the relationship between copper and silica contents of the Hindubagh rocks; it shows clearly that copper

tends to be concentrated in the earlier formed fractions.

NICKEL: Several investigators, *e.g.*, Vogt (1923), Sandell and Goldich (1943), Wager and Mitchell (1951), Nockolds and Allen (1956), Storm and Holland (1957), and others, have shown that Ni is concentrated in the early differentiates, a finding that the present investigation confirms. The highest concentration (1227 ppm) in the Hindubagh rocks is in a dunite specimen that has a MgO content of (MgO, 49.52 per cent) show up to a maximum of 0.64 per cent NiO and confirm the suggestion.

The distribution of Ni in the Hindubagh rocks shows close similarity with that in tholeiitic basalts of Mauna Loa and Kilauea (Nockolds and Allen, 1956, p. 71, Tables 28, 29). A comparison of the distribution of Ni in the Hindubagh rocks with those of other areas shows that Ni content of these rocks is normal (Table 4).

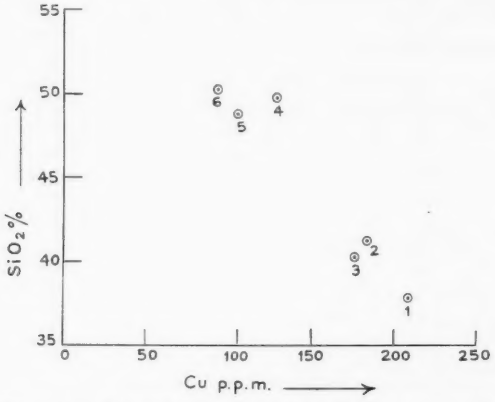

Figure 3. Relationship between Cu and SiO_2 contents of Hindubagh rocks

43.42 per cent. Figure 4 shows that Ni follows MgO closely in this suite. Nockolds and Mitchell (1948) and Ringwood (1955) have criticized the commonly held view that Mg camouflages Ni. Ringwood thinks it unlikely that Ni can proxy extensively for Mg and suggests that it enters crystal lattice at the expense of Fe^{+2}. Nockolds and Mitchell (1948) have suggested that concentration of Ni in the early differentiates may be due partly to the replacement of Fe^{+2} by Ni and partly to the greater polarizing power of Ni, which results in its behaving as if it had a smaller ionic radius.

Sandell and Goldich (1943, p. 175) have suggested that olivine is the most important Ni-bearing mineral among the silicates and that NiO content increases with the increase in Mg_2SiO_4 molecule in olivines. The 13 analyses of olivines given by Goldschmidt (1923, p. 307) and 2 given by Sandell and Goldich (1943, Table 12) confirm this. Hindubagh olivines

The Ni–Co ratio in the Hindubagh rocks ranges from 11:1 for sample 1 to 2:1 for sample 6. This appears to be the normal trend and is also exhibited by the Skaergaard rocks (Wager and Mitchell, 1951, p. 189), the Karroo dolerites, and the British tholeiitic series (Nockolds and Allen, 1956, p. 46, 51, 53, Figs. 43c, 43e). This marked change in the Ni–Co ratio appears to be due partly to the closer ionic relationship between Mg and Ni and Fe^{+2} and Co and partly to the fact that Ni is taken up early in the differentiation, whereas Co does not exhibit any tendency for early concentration. Lundegårdh (1946b, p. 159–161) has suggested that Ni–Co ratio indicates the stage of differentiation reached in a rock series. Wager and Mitchell (1951, p. 189) confirm this suggestion for mafic and intermediate rocks of the Skaergaard intrusion.

COBALT: Various investigators disagree on the behavior of Co in igneous rocks. Sandell and Goldich (1943, p. 138) have shown that the re-

lationship between Co and Mg is more apparent than that between Co and Fe^{+2}. Nockolds and Mitchell (1948) have, however, pointed out that since ionic radii of Co^{+2} and Fe^{+2} (0.72 Å and 0.74 Å) are very close, Co enters the crystal lattice with the same facility as Fe^{+2}. In the present study Co does not show any systematic variation either with respect to Mg or Fe^{+2}, and this would appear to suggest that Co can proxy for Mg as well as Fe^{+2} depending upon the conditions prevailing at the time of crystallization. The distribution of Co in the Hindubagh rocks is similar to that exhibited by Hawaiian tholeiitic basalts of Mauna Loa and Kilauea (Nockolds and Allen, 1956, p. 48). The over-all distribution of Co in the Hindubagh rocks is close to the average for mafic rocks given by Goldschmidt but is lower for ultramafic rocks (Table 4).

VANADIUM: Goldschmidt (1954, p. 489) and Wager and Mitchell (1951, p. 185) have shown that vanadium follows Fe^{+3} closely. They have explained this as being due mainly to the similar dimensions of the ionic radii of the two elements (V^{+3} = 0.65 Å, Fe^{+3} = 0.67 A). Ahrens (1952, p. 168), however, gives 0.74 A as the ionic radius of V^{+3}, which is closer to the value of Fe^{+2} (0.74 Å). It has, however, become clear from various trace-element studies that V^{+3} does not follow any major element. In the Hindubagh rocks, as in the Skaergaard intrusion (Wager and Mitchell, 1951, p. 185), vanadium is concentrated in the intermediate rocks, although the comparatively high concentration (133 ppm) of the element in specimen 2 (Table 5) may be due to its entry in the spinel, which constitutes 5.5 per cent of the rock. The distribution of vanadium is remarkably uniform in the Hindubagh rocks, and the trend is closely similar to that of the Skaergaard complex. The concentration rises from 93 ppm to 133 ppm and then falls to 68 ppm. Rocks of Hakone P-series (Nockolds and Allen, 1956, p. 47, Fig. 43a) show the same trend. This,

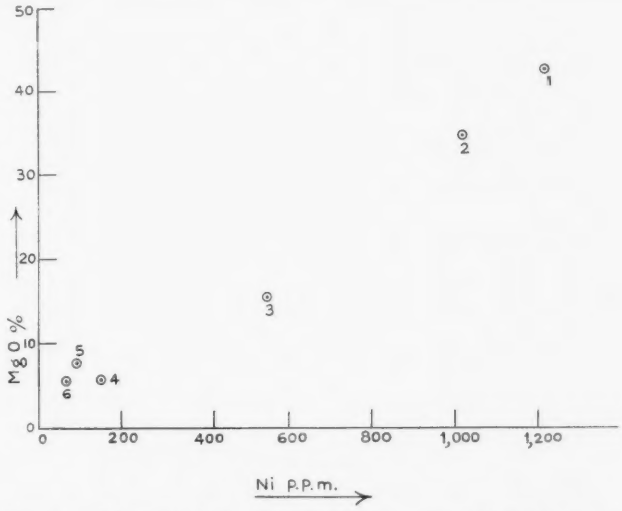

Figure 4. Relationship between Ni and MgO contents of Hindubagh rocks

TABLE 4. DISTRIBUTION OF Cu, Ni, Co, V, AND Cr IN ULTRAMAFIC, MAFIC, AND INTERMEDIATE ROCKS (PPM)

	Ultramafic	Mafic	Intermediate
Cu*	..	149	38
Ni†	3600	160	40
Co†	240	80	31
Co**	90	50	20
V††	130	250	130
Cr†	5000	500	100

* Sandell and Goldich (1943)
† Goldschmidt (1954)
** Wager and Mitchell (1951)
†† Troger (1935)

however, is not the only trend; Nockolds and Allen (1956, p. 47, Figs. 43b, 43a) have demonstrated that in the rocks of the Hakone H-series and Karoro, vanadium is concentrated in the early fractions. More work on the geochemistry of vanadium is required before its distribution in the igneous rocks can be fully explained.

TABLE 5. DISTRIBUTION OF Cu, Ni, Co, V, AND Cr IN ROCKS OF HINDUBAGH IGNEOUS COMPLEX (PPM)

Analyst S. A. Bilgrami

	1	2	3	4	5	6
Cu						
C.D.	200	190	178	130	100	90
	220	185	170	125	105	91
S.D.	205	180	182	132	107	95
Average	208	184	176	129	104	92
Ni						
C.D.	1210	1075	565	150	92	66
	1240	1035	545	148	93	67
S.D.	1230	1050	530	150	88	72
Average	1227	1053	547	149	91	68
Co						
C.D.	118	98	112	50	44	35
	115	105	100	52	46	35
S.D.	115	100	110	49	47	33
Average	116	101	107	50	46	34
V						
C.D.	90	135	115	65	90	85
	95	130	105	70	84	95
S.D.	95	135	110	70	82	88
Average	93	133	110	68	85	89
Cr						
C.D.	3625	650	135	105	290	85
	3405	680	150	95	270	75
S.D.	3200	640	145	90	258	78
Average	3410	657	143	97	276	79
Ni/Co	11:1	10:1	5:1	3:1	2:1	2:1

C.D. = colorimetric determinations; S.D. = spectrochemical determinations. Numbers in column headings explained in footnote to Table 2

CHROMIUM: Goldschmidt (1954, p. 548) has shown that concentration of chromium in igneous rocks is related to the sequence of crystallization, the earlier formed rocks being richer in chromium than the later ones. The investigations of Nockolds and Mitchell (1948) and Wager and Mitchell (1951) and the present investigation have confirmed this. The range of chromium content in the ultramafic and mafic rocks given by Goldschmidt (1954, p. 548) is 1000–44000 and 100–400 ppm respectively, and in this respect the Hindubagh rocks do not show any abnormality. In the dunite specimen (Table 5, analysis 1) most of the chromium is located in the spinel chromite; the same applies to analysis 2, although in this case the spinel is a golden-brown picotite. The distribution of chromium in the Hindubagh rocks appears to be related in a general way to MgO, except in specimen 312 (Table 5, analysis 5), which is high in ferromagnesian minerals, and it is likely that most of the chromium is located in the silicate minerals, probably replacing Fe^{+3}. Wager and Mitchell (1951, p. 183) have shown that the concentration of chromium in the Skaergaard pyroxenes is very high (3000 ppm), and this would seem to justify the assumption that most of the 276 ppm of chromium in specimen 312 is in the ferromagnesian minerals.

Vogt (1921, p. 323) has suggested that if the concentration of chromium in a magma is 1000 ppm, chromite will form as a separate phase. Wager and Mitchell (1951) have shown that in the case of the Skaergaard intrusion an opaque chrome spinel separated out even though the chromium content of the magma was only 170 ppm. The numerous chromite veins in the Hindubagh area suggest that the concentration of chromium in the original Hindubagh magma was fairly high. The abrupt fall in the chromium content of the successive rocks clearly shows that the chromite crystallized during a relatively short time. Petrological evidence confirms this. Chromite is not present in any of the mafic rocks of the complex.

CONCLUSIONS

The distribution of trace elements in the rocks of the Hindubagh igneous complex is normal. It is, however, obvious that variations will be discovered as more work is done on the complex.

The colorimetric and spectrochemical determinations of the trace elements (Table 5) agree closely, except in the case of chromium which persistently gave low spectrochemical results. Possibly the current used in arcing was not high enough. It would not be safe to conclude from this study that one method of trace-element determination is as good as the other since the magnitude of distribution of trace elements was known in this case before the spectrochemical work was undertaken. This enabled the preparation of more closely calibrated reference samples than would have been possible had the concentration not been known.

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Late Quaternary History of the Snake River in the American Falls Region, Idaho

Abstract: While mapping the American Falls region, we found evidence that contributes to the middle Pleistocene to Recent history of the Snake River, and indirectly to the history of overflow of Lake Bonneville. Middle Pleistocene to recent rocks in the valley are mainly lacustrine and fluvial silts and clays, with some sand, gravel, basalt, and a few thin tuff beds. The formation of terraces can be correlated with events both up- and downstream.

The Snake River was at least once, and possibly twice, dammed and diverted by eruptions of basalt, resulting in the formation of lakes and deposition of lacustrine beds. A rather flat-lying, thin, but persistent gravel at the base of one lake bed formation may represent a glacial period, possibly Illinoian, during which the Snake River had a large volume.

Overflow of water from Lake Bonneville into the Snake River system, by way of the Marsh Creek-Portneuf valley, laid down a deltaic-fluvial deposit here named the Michaud Gravel. At this time the Snake River, greatly augmented by Lake Bonneville overflow, began to cut channels through and around a lava dam. Terraces between Aberdeen, American Falls, and Pocatello were formed during the existence of the lake in which the Michaud Gravel was deposited and by fluvial processes after drainage of the lake. At one stage in the downcutting, bars of huge basalt boulders were built across the mouths of abandoned spillways.

Radiocarbon dating and geologic evidence from the area between Preston and Soda Springs, Idaho, suggest that basalt flows diverted the Bear River into Lake Bonneville, perhaps causing it to overflow. This diversion probably occurred about 33,000 years ago. This dating accords with events in the American Falls region.

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INTRODUCTION

Detailed geologic mapping of the American Falls and Rockland 15-minute quadrangles, Idaho, and reconnaissance of the Yale and Michaud 15-minute quadrangles and adjacent areas by the U. S. Geological Survey (Fig. 1) has revealed previously undescribed events in the history of the Snake River valley and the relation of these events to the spillover of Lake Bonneville at Red Rock Gap. In the American Falls region the late Quaternary history can be closely related to the local stratigraphy.

Robert C. Bright contributed valuable evidence from the Preston, Idaho, area.

The area is in southeastern Idaho (Fig. 1) along the Snake River valley from Pocatello southwestward to the mouth of the Raft River. The Snake River valley in this region consists of (1) a relatively flat wide upstream area around the American Falls Reservoir in which the Snake River has a low gradient, (2) a

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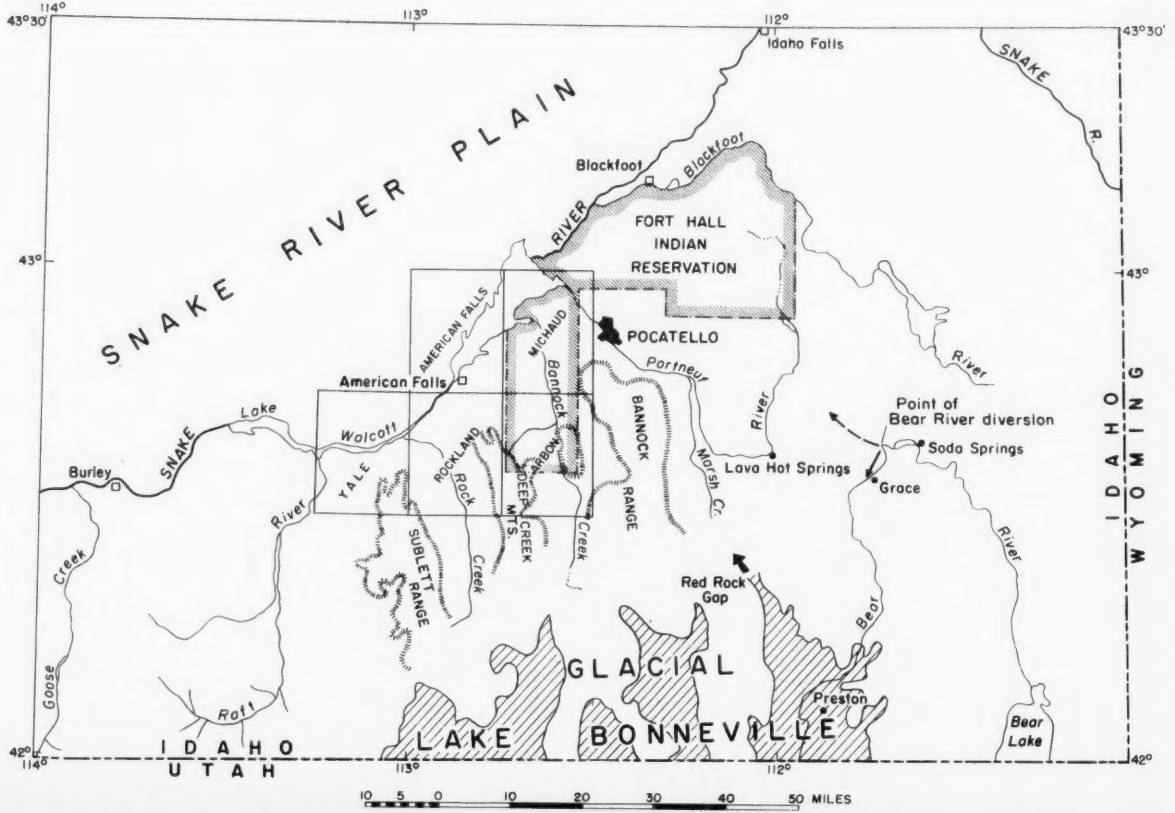

Figure 1. Index map of southeastern Idaho showing the American Falls region, its relation to glacial Lake Bonneville, and other geographic features

middle area below American Falls, where the river has a gradient of 7 feet per mile, and through which it has carved a canyon 50 to 150 feet deep, partly in Recent time, and (3) a downstream area below Massacre Rocks, now occupied by Lake Walcott and impounded by Minidoka Dam, where the gradient is low, but where part of the channel passes through a shallow canyon, cut mainly in late Pleistocene time. The upstream area is underlain by soft sediments with some intercalated basalt on the northwest side of the valley. The middle part of the valley is cut mainly in older basaltic and rhyolitic rocks, and the lower part is cut along the boundary between basalt and weakly consolidated sedimentary deposits.

G. K. Gilbert (1890) was the first to report in detail upon the Quaternary history of the Bonneville basin, but he did not study the Snake River valley itself. He recognized (p. 60) the point of spillover of Lake Bonneville at Red Rock Pass and its significance in relation to lake stages. He also recognized three important high Lake Bonneville stages, the oldest of which he called the "Intermediate stage" (p. 135–154) because its deposits lie between the Bonneville and Provo levels in elevation. The Provo is the youngest of the three stages. Part of the deposits of Gilbert's "Intermediate stage" were later named the Alpine Formation by Hunt and others (1953, p. 17).

The first important geologic study of the Snake River Plains was by Russell (1902). Mansfield, during his study of the Fort Hall Indian Reservation, made the first detailed geologic map (1920, Pl. III) in the American Falls region. He did not examine the deposits of the Snake River valley in detail but did recognize three erosion cycles, the younger two of which are partly equivalent to fluvial terraces mapped by the writers. Later Mansfield (1927, p. 15–20) proposed a more complex series of erosion cycles for southeastern Idaho.

Stearns and others (1938) made the only previous comprehensive geologic study of the eastern Snake River valley, but much of the report concerns ground-water resources of the region. They named most of the Tertiary formations and noted (p. 89) that, "essentially all the [alluvial] material [along the Snake River] was laid down either in impounded waters behind lava dams, or as fans at the toes of these dams." They emphasized the role of springs along the Snake River as agents of erosion but failed to recognize the importance and timing of Lake Bonneville spillover in the late Pleistocene history of the Snake River valley. More recently Stearns and Isotoff (1956) mapped the Massacre Rocks-Eagle Rock area. Poulsen, Nelson, and Poulsen (1943) studied the soils of part of the area and named several terraces.

Modern contributions to the study of Lake Bonneville have been numerous, and much discussion of the chronology and stratigraphy of the various lake stages has been published (Bissell, 1952; Richmond and others, 1952; Williams, 1952; Hunt and others, 1953; Marsell and Jones, 1955; Eardley and others, 1957; Feth and Rubin, 1957; Broecker and Orr, 1958). These papers indicate many problems and lack of agreement on the chronology of Lake Bonneville, much of which stems from differences in radiocarbon age determinations on marl and tufa from the lake.

The importance of the Lake Bonneville spillover in the late Quaternary history of the Snake River valley was first noted by H. A. Powers and H. E. Malde in the western Snake River Plain. They interpreted boulder bars and abandoned cataracts as caused by torrential waters added to the river by Lake Bonneville spillover (Malde, 1960).

STRATIGRAPHY

Rocks exposed along the Snake River include rhyolitic and welded tuff of Pliocene age; basalt and basaltic tuff of probable early Pleistocene age; clay, silt, sand, and gravel of middle to late Pleistocene age, mostly lacustrine and fluvial; basalt of middle to late Pleistocene age; and alluvium and dune sand of Recent age.

The oldest rocks prominently exposed in the Snake River canyon are the Pliocene Neeley Lake Beds (Stearns and others, 1938, p. 43) and the Walcott Tuff of Stearns (Stearns and Isotoff, 1956, p. 23). These two conformable formations crop out mainly along the south side of the Snake River from American Falls southwestward to Rock Creek. The Neeley consists of about 100 feet of poorly bedded light-tan to orange-brown fine- to coarse-grained rhyolitic tuff, with lenses of pebbles and pumice fragments. The overlying Walcott is about 15 to 75 feet thick and consists of a lower white bedded tuff unit and an upper welded obsidian tuff unit.

Unconformably above the Walcott Tuff, and having about the same distribution, is an unnamed unit, 10 to 100 feet thick, of probable early Pleistocene age, consisting of basaltic tuff

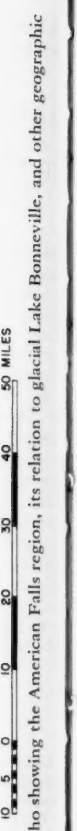

Figure 1. Index map of southeastern Idaho showing the American Falls region, its relation to glacial Lake Bonneville, and other geographic features

with rhyolitic pyroclastic debris, overlain unconformably by basalt. Stearns *et al.* (1938, p. 47) included this formation in the Massacre Volcanics, of which the basalt is a part.

The next youngest formation, and the oldest with which this paper is directly concerned, is named the Raft Formation. This unit formerly was called the Raft Lake Beds (Stearns and others, 1938, p. 48). The name is changed because much of the unit, as indicated by Stearns and others (p. 40, 50) and as observed by us, appears to be fluviatile rather than lacustrine, although much of the formation probably did accumulate in a lake. The Raft Formation includes part of what Stearns *et al.* (p. 69) called the American Falls Lake Beds. The lower part of his American Falls consisted of beds older than the basalt that dammed the Snake River and created the lake in which the American Falls Lake Beds were deposited. Thus, in the bluffs on the Snake River opposite Neeley, we assign to the Raft approximately the lower two-thirds of the exposed beds. The Raft Formation is exposed at scattered places along both sides of the Snake River below American Falls, but chiefly in the bluffs opposite Neeley and below Rock Creek on the south side of the river (Pl. 1). The Raft is mainly light-colored poorly bedded to nearly massive silt in the upper part and tan and gray clay, silt, and fine sand in the lower part. It contains a few gravel lenses, many irregular nodules and persistent layers of calcium carbonate, and a few local layers of basaltic and rhyolitic tuff. The Raft ranges from a few feet to more than 200 feet thick. Originally considered Pliocene(?) by Stearns and others (1938, p. 48), it is now known to be Pleistocene on the basis of mollusk collections.

Above the Raft, but possibly intertonguing with it, is a widespread group of nearly contemporaneous basalt flows from several vents northwest of the Snake River. These flows are restricted to the northwest side of the Snake River valley. They contain conspicuous olivine crystals, much magnetite, and glass in the groundmass, and are normally vesicular and fresh looking. These basalts are part of what has been mapped as Snake River Basalt (Ross and Forrester, 1947) and "undifferentiated basalt" (Stearns and others, 1938, Pl. 4).

Cedar Butte Basalt (Stearns and others, 1938, p. 69), which is younger than most of the Snake River Basalt except for flows of Recent age outside the area, occurs in an area about 8 or 10 miles long by 4 or 5 miles wide, almost entirely west of the present Snake River canyon. It probably overlies the Raft Formation in most places, but exposures of this contact are poor. The Cedar Butte issued from vents about 11 miles southwest of American Falls. Over most of the area it is a single flow about 25 to 75 feet thick but locally it consists of at least two olivine basalt flows totaling 200 feet.

The American Falls Lake Beds, which were deposited immediately after emplacement of the Cedar Butte Basalt, overlie the Snake River Basalt or the Raft Formation, where the basalt is absent. The lower part of the American Falls Lake Beds of Stearns and others (1938, p. 69) is now included in the Raft Formation. The lake beds are exposed along the northwest side of the Snake River from a point 1 mile northeast of Eagle Rock to the upper end of the American Falls Reservoir. They are 40–80 feet thick and consist of light-colored bedded silt, sand, clay, and one persistent layer of blocky white diatomaceous clay. A layer of peat occurs near the top of the formation 5 miles northwest of Pocatello. At the base of the formation a thin, discontinuous but persistent layer of gravel is included, although this gravel is probably not related to the lake beds.

Above the American Falls Lake Beds are deltaic sand and gravel deposits, and several terrace deposits of silt, sand, and gravel, all mainly north and east of American Falls. The deltaic sand and gravel, which marks an important event in the Pleistocene history, is named the Michaud Gravel. The type locality is in gravel pits on the south side of the Union Pacific Railroad at Michaud, Idaho, about 8 miles west of Pocatello. The Michaud is composed of sand and gravel that contains boulders as much as 8 feet in diameter at Pocatello, but the size diminishes rather abruptly westward to small cobbles at Michaud, and to a sand with a few small pebbles northeast of American Falls. The pebbles are mainly quartzite of Paleozoic age, but limestone and volcanic rocks are also present. The Michaud is about 5 to 50 feet thick.

Three terraces are distinguished between American Falls and Aberdeen (Pl. 1) (from oldest to youngest): the Grandview, Aberdeen, and Sterling terraces, named by Poulsen, Nelson, and Poulsen (1943, p. 5). The terraces are covered by 5 to 25 feet of light-colored alluvium of sand, silt, and clay, and, on the Sterling and Aberdeen terraces, local gravel. These deposits are delineated mainly by topographic position; the Grandview lies be-

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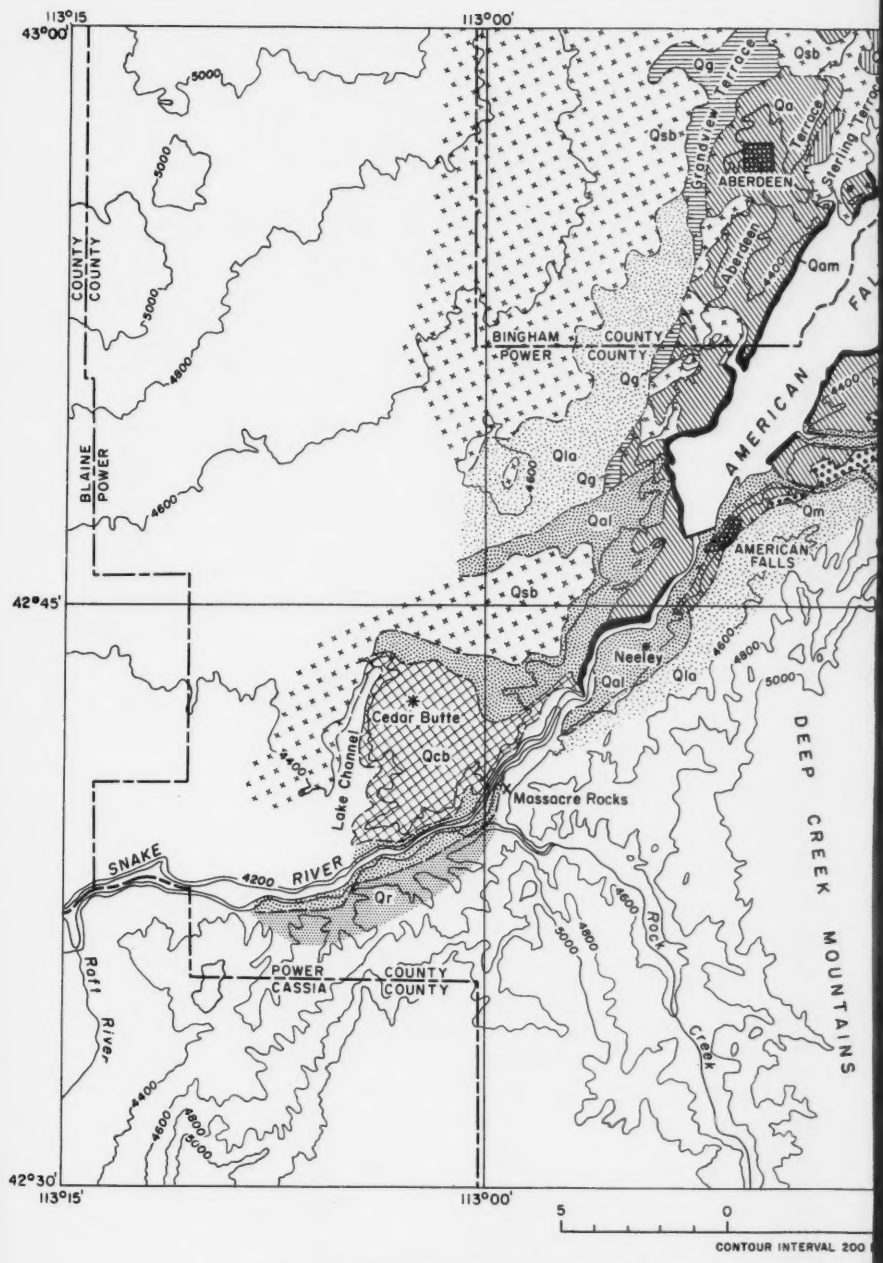

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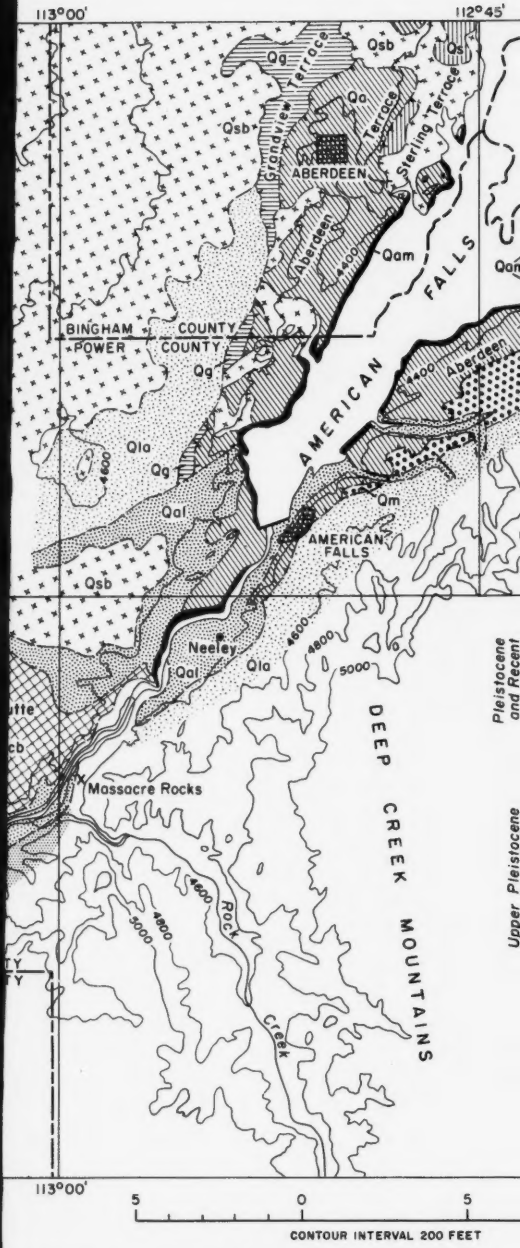
TRIMBLE AND CARR, PLATE 1
Geological Society of America Bulletin, volume 72



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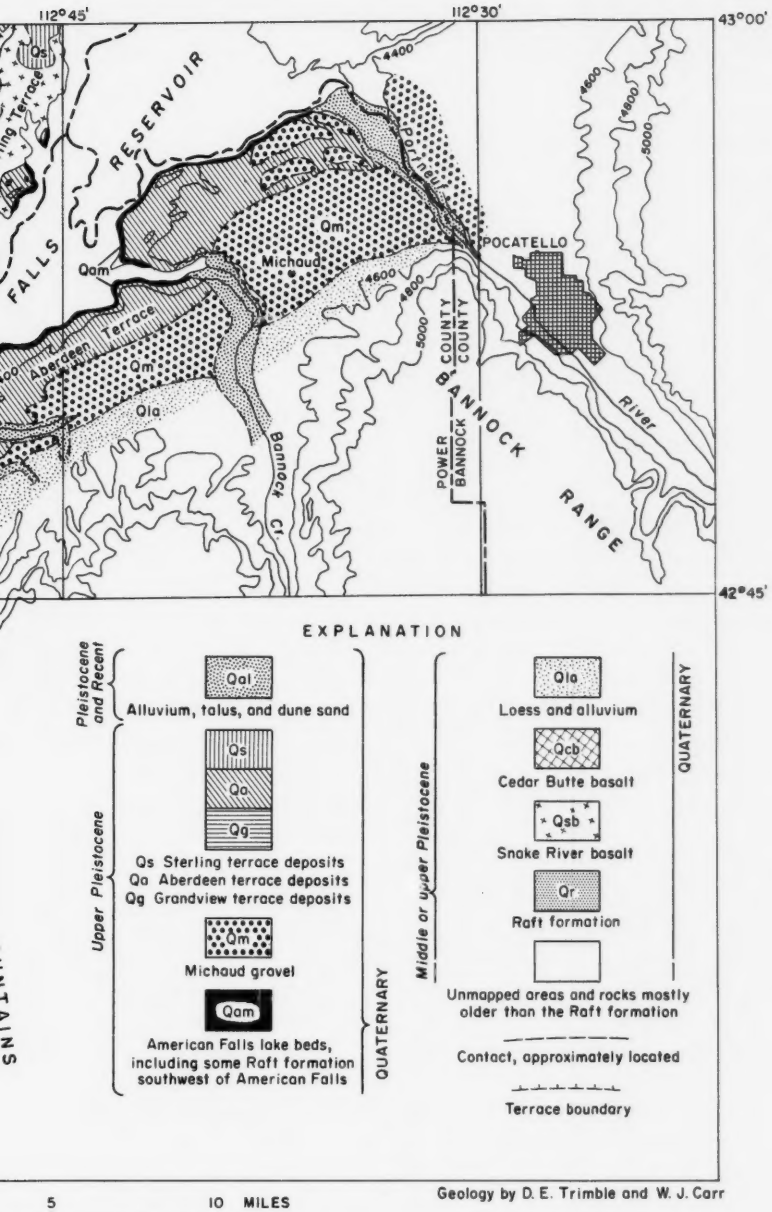

Geology by D. E. Trimble and W. J. Carr

ALLS REGION, IDAHO, SHOWING QUATERNARY ROCKS

tween altitudes of about 4430 and 4460 feet, the Aberdeen between 4390 and 4430 feet, and the Sterling between 4360 and 4390 feet.

The Aberdeen terrace is cut partly on Michaud Gravel and is therefore younger than the Michaud.

Latest Pleistocene and Recent deposits in the area consist of alluvium, dune sand, and talus. Most of the older alluvium is reworked from, and is lithologically indistinguishable from parts of the Michaud Gravel and Aberdeen terrace deposits. It is distributed from nearly present river level up to about 4460 feet in altitude, mainly southwest of American Falls. Sand dunes of Recent age occur in a large northeast-trending belt paralleling the Snake River along its north side. The dunes are locally active, and most trend northeasterly. Recent alluvium is restricted to a few gravel bars along the Snake River and some fine-grained sediments in the American Falls Reservoir area.

GEOLOGIC HISTORY

Deposition of the Raft Formation and Overlying Gravel

In middle or late Pleistocene time the Raft Formation was deposited, largely in a lake formed by damming of the river channel by basalt flows, possibly immediately west of the Raft River. Lake beds in the Raft require damming to an altitude of about 4400 feet. A persistent zone of marlstone occurs at an altitude of about 4280 feet and rises eastward to 4350 feet, representing a westerly dip of about 10 feet per mile. This dip may be due to regional tilting. On the margins of the depositional basin of the Raft Formation southwest of Rock Creek, local gravelly and sandy beds were laid down, generally above 4400 feet. A few beds of basaltic tuff and rhyolitic tuff in the Raft indicate that volcanic eruptions continued but were much less voluminous.

After deposition of the Raft Formation, through drainage was re-established, and a very gentle valley with its axis nearly in the present position of the Snake River formed in the Raft. Local base level was about 4320 feet, and the gradient of the ancestral Snake River was probably less than 1 foot per mile. Wells about 3 and 6 miles northwest of American Falls encountered the top of the Raft at about 4340 and 4350 feet, respectively, indicating a slope of about 5 feet per mile on the sides of the valley.

A mixture of rhyolitic and basaltic ash and a thin but widely persistent pebbly sand and gravel were deposited on this very gentle slope. Most of the pebbles are less than 1 inch in diameter, and many are from distant sources. It is difficult to explain the transportation of this gravel in such a gentle valley. Most of the pebbles could not come from the Deep Creek Mountains to the south, nor is any local older gravel known from which they could have been derived. A constriction of channel or considerable increase in volume and velocity of the river, seems to be probable. The Snake River Basalt, some of which probably was erupted about this time, could have reduced the width of the valley from about 5 to 1.5 miles in the American Falls area. This, however, would have been insufficient to account for the increase in velocity of the river, so we conclude that temporarily the Snake carried much more than normal volume. At least 80 per cent of the pebbles in this deposit are pink Brigham Quartzite of Cambrian age, which could have been derived only from the three major drainages at the east side of the area—Bannock Creek, the Portneuf River, and Ross Fork Creek. Very few basalt pebbles are present in the gravel. Presumably a substantial portion of the water responsible for moving this gravel came down these streams.

The gravel overlying the Raft Formation has yielded bones of birds, mammoths, horses, camels, bison, and other animals, as well as fresh-water clams and snails, and land snails. Skulls of the giant bison, *Bison* (*Gigantobison*) *latifrons* (Harlan), were described by Hopkins (1951, p. 192–197). These specimens came from the gravel below the American Falls Lake Beds and above the Raft Formation, at the east edge of the American Falls Reservoir. Hibbard (1955, p. 221) believes that *Bison latifrons* is no older than Illinoian age. However, Schultz and Frankforter (1946, p. 8) consider the giant bison to be Kansan and Yarmouth. From the same gravel at another locality we collected bones of mammoth or mastodon, genus indeterminable, *Equus* sp., *Camelops* sp., and *Bison* sp. These were identified by G. E. Lewis, C. B. Schultz, and L. G. Tanner (Written communication, 1958), who suggested an age between late Kansan and middle Illinoian and compared the association of mammoth, horse, and camel to a similar association in the Sappa Formation of Nebraska, of Yarmouth age (Schultz and Tanner, 1957, p. 64, 66, 70, 74). The Sappa is latest Kansan or early Yarmouth

(Condra and others, 1950, p. 22; Frye and others, 1948, p. 520).

D. W. Taylor believes that mollusks from the underlying Raft Formation are middle or probably late Pleistocene (Written communication, 1959). The following species from the Raft and the overlying gravel identified by Taylor, suggest a cool wet environment.

	Raft Formation	Gravel at base of American Falls Lake Beds
Fresh-water clams		
Sphaerium striatinum (Lamarck)	×	×
Pisidium compressum Prime	×	×
Fresh-water snails		
Valvata humeralis (Say)	×	×
V. utahensis Call	×	×
Lithoglyphus fuscus (Haldeman)	..	×
Stagnicola caperata (Say)	×	?
S. palustris (Müller)	..	×
S. traskii (Tryon)	×	..
S. sp. a	×	..
S. sp. b	×	..
Gyraulus circumstriatus (Tryon)	..	×
G. parvus (Say)	×	×
Carinifex newberryi (Lea)	..	×
Helisoma anceps (Menke)	..	×
Promenetus exacuous (Say)	..	×
P. umbilicatellus (Cockerell)	..	×
Ferissia indet.	..	×
Physa indet.	..	×
Land snails		
Pupilla muscorum (Linnaeus)	..	×
P. indet.	×	..
Vertigo gouldi (Binney)	..	×
Vallonia cyclophorella (Sterki)	×	×
cf. *Succinea*	×	×
Discus cronkhitei (Newcomb)	..	×
D. shimeki cockerelli Pilsbry	×	..
D. indet.	×	..
Retinella indet.	..	×
Hawaiia minuscula (Binney)	..	×
Zonitoides arboreus (Say)	..	×
Oreohelix strigosa depressa (Cockerell)	×	..

Most of the above mollusks also occur in the overlying American Falls Lake Beds, which Taylor (written communication, 1959) believes are late Pleistocene.

The gravel is almost certainly late Kansan or early Wisconsin. We believe that the increased stream velocity, indicated by the ability to move gravel on a very gentle slope, and the cool moist climate suggested by the fossils imply a glacial rather than an interglacial environment of deposition. When the age of *Bison latifrons* has been determined, the gravel can be dated with more confidence.

Damming of the Snake River and Deposition of the American Falls Lake Beds

After deposition of the gravel above the Raft Formation and extrusion of Snake River Basalt, eruptions from Cedar Butte and associated vents 11 miles southwest of American Falls filled at least 8 miles of the Snake River valley with basalt and dammed the river (Fig. 2). The Cedar Butte Basalt abutted the Snake River Basalt on the north and the Walcott Tuff and Massacre Volcanics on the south, forming a large dam. The upper surface of the dam sloped gently upstream from an altitude of more than 4450 feet near the vents to about 4370 feet at the upstream edge. Water was impounded in a lake that extended upstream about 40 miles nearly to Blackfoot. The maximum altitude of its surface was about 4450 feet, the approximate minimum altitude of the Cedar Butte Basalt dam at its northern and southern margins and the altitude of the base of the highest wave-cut scarp marginal to the American Falls lake. Overflow poured around the margins of the dam through two major spillways, one discharging in about the present position of Lake Channel, the other along the present Snake River channel.

The American Falls Lake Beds were deposited in this lake and accumulated to an altitude of about 4400 feet. On the west side of the lake they were deposited on the fringes of Snake River Basalt. Elsewhere they rested mainly on the Raft Formation and the gravel deposit. At first mostly fine sand and silt were deposited in the lake, except eastward from the mouth of Bannock Creek where Bannock Creek and the Portneuf River built deltas of crossbedded sand. Conditions then changed, and 5 to 15 feet of white, blocky, very diatomaceous clay was deposited throughout the lake basin. In the eastern end of the lake, the clay was laid down on an undulating surface cut by streams in the deltaic deposits.

Overflow of Lake Bonneville

When the American Falls lake had become two-thirds filled with sediments, glacial Lake Bonneville overflowed through Red Rock Gap about 50 miles southeast of Pocatello. The water descended northward by way of Marsh Creek and the Portneuf River valleys, entered the shallow American Falls lake near Pocatello, and deposited the Michaud Gravel on the lake

beds. As waters entered the lake much of the tremendous energy of this flood abated, and huge well-rounded boulders as much as 8 feet in diameter were deposited at Pocatello. Pebbles and cobbles were carried some distance into the lake.

the site of a receding falls approximately along the present Snake River channel and was cut mainly along the southern margin of the Cedar Butte Basalt. As the heads of the two channels receded upstream, the altitude of the head of the southern channel became lower more

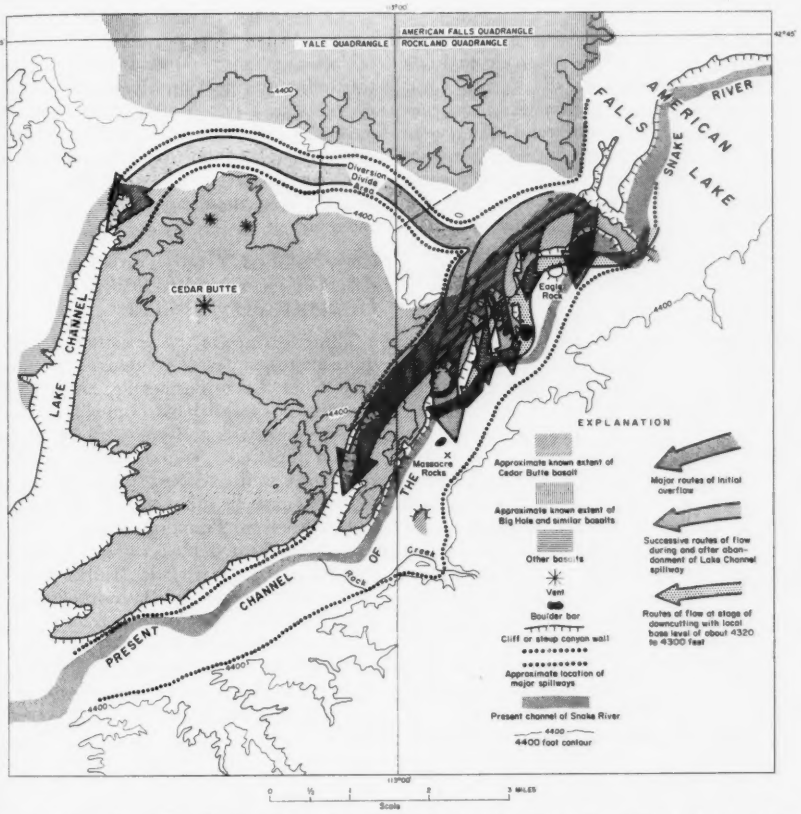

Figure 2. Map of area of spillover of American Falls lake showing Cedar Butte Basalt dam and routes of Lake Bonneville-Snake River flow

The tremendous volume of Lake Bonneville overflow added to the Snake River drainage, greatly increased the discharge from the American Falls lake, and accelerated the cutting of the two major spillways (Fig. 2) around the Cedar Butte Basalt dam. Headward cutting of the northern spillway resulted in a falls at the head of a basalt-rimmed canyon, now called Lake Channel. The southern spillway was also rapidly than the northern stream could erode its course, so that, when the head of the southern channel fell to an altitude of about 4400 feet, all the water was diverted into the southern spillway.

Just before this diversion a persistent scarp was cut at an altitude of about 4420 feet on beds marginal to the American Falls Lake Beds and on the Michaud Gravel. This scarp marks

the front of the Grandview terrace and the back of the Aberdeen terrace. Pebbles of Paleozoic rocks on the floor of the abandoned channel indicate fluvial conditions upstream. Sand and some gravel were deposited on the fore part of the present Aberdeen terrace. The lack of coarse material suggests cessation of initial outflow of Lake Bonneville at this time.

When the Lake Channel spillway was abandoned, the falls in the other channel was probably between Massacre Rocks and Eagle Rock (Fig. 2), about 2 or 3 miles downstream from the eastern margin of the Cedar Butte Basalt. Marginal spillways were successively cut and abandoned as the falls in the main channel receded upstream across a gradually lower surface of the Cedar Butte Basalt. The base level of cutting downstream from the falls was about 4320 feet or about 120 feet above present river level.

The fluvial Sterling terrace, with a back at an altitude of about 4380 feet, was cut at this time. It probably was graded temporarily to the Cedar Butte Basalt dam near its eastern margin, or possibly to the basalt at American Falls.

When the river finally breached the dam about a mile northeast of Eagle Rock, rapid downcutting occurred upstream in the soft sediments. Bars of huge, locally derived basalt boulders (mainly Cedar Butte Basalt), some as much as 20 feet in diameter, were built across the mouths of abandoned spillways between Eagle Rock and Massacre Rocks. Some of these bars may have been formed by water occupying the marginal spillway. Others, however, cannot be explained so simply, and their location suggests that they may have been emplaced in part because of the confinement of a large volume of overflow waters from Lake Bonneville to a narrower channel at and below the former dam with resulting increased velocity and competency of the stream.

Return of Normal Snake River Flow

Overflow of Lake Bonneville ceased, and in latest Pleistocene and Recent time the river lowered its channel about 100 feet in the Eagle Rock area. In upstream areas it has reworked downward the gravels that once belonged to the Michaud Gravel and Aberdeen terrace deposits. More recently, the river has deposited a little gravel as bars in the channel. Some of the reworked gravel, however, may be intermediate in age between the Aberdeen terrace deposits and Recent alluvium, possibly correlating with the Sterling terrace deposits upstream. One deposit of older alluvium is the gravel at the east end of the American Falls Dam, from which Gazin (1935) collected a large and varied vertebrate fauna from an undetermined part of the Pleistocene. This fauna includes musk ox, considered indicative of glacial periods; the deposit probably is Wisconsin.

Upstream in the American Falls Reservoir area, alluvium, now mostly inundated by the reservoir, was deposited in the valley. Currently active longitudinal sand dunes have developed along the northwest side of the valley. Construction of the American Falls Dam has resulted in accumulation of fine sediments in the reservoir.

CHRONOLOGY OF EVENTS IN RELATION TO SPILLOVER OF LAKE BONNEVILLE

Three features of the area that seem to require abnormally large volumes of water, probably from Lake Bonneville, are (1) large boulders in the Michaud Gravel at Pocatello; (2) the northern and southern channelways around the Cedar Butte Basalt dam, which apparently were occupied simultaneously; and (3) the boulder bars in the Snake River canyon.

The Michaud Gravel is older than 30,000 years, as indicated by a radiocarbon date of 29,700 ± 1000 years B.P. (Rubin and Alexander, 1960, sample W-731) for shells from the deposits of the Aberdeen terrace, which is younger than the Michaud Gravel.

The effects of Lake Bonneville spillover seen at American Falls are probably related to drainage changes in the Bear River studied by R. C. Bright. The age of shells from high shore-line deposits of a former lake near Thatcher, Idaho, has been determined as 32,500 ± 1000 years B.P. (Rubin and Alexander, 1960, sample W-704). This Thatcher Lake was formed by diversion of the Bear River, and the lake spilled into the Bonneville basin. Bright has suggested that this diversion of Bear River into the Bonneville basin may have caused overflow at Red Rock Pass (Rubin and Alexander, 1960). On the basis of the inferred history at American Falls the Michaud Gravel is not much older than the Aberdeen terrace, and could be the same age as the diversion of the Bear River.

Bright (written communication, 1961) has also recognized in the Thatcher area a shore

line of the Bonneville stage of Lake Bonneville, which is much younger than the shore lines of the earlier Thatcher lake. The deposits of the older lake were excavated nearly to the level of the present valley floor before the Bonneville stage deposits were superimposed. This lowering of base level requires that a divide between Thatcher and Preston, Idaho, be cut down about 700 feet, mostly through quartzite. The Bonneville stage deposits in the Thatcher area have not yet been dated, but they are obviously much younger than the older Thatcher Lake. The Michaud Gravel, therefore, which probably is contemporaneous at least in part with the older lake, is considerably older than the Bonneville stage.

The channels around the Cedar Butte Basalt dam were almost certainly cut by water of the same stage of spillover that deposited the Michaud Gravel. The boulder bars, too, might possibly be attributed to this stage. It is conceivable, therefore, that all the features of this area that indicate a large volume of water were formed during a single episode of overflow of Lake Bonneville older than Bonneville stage.

The fineness of the Aberdeen and subsequent Sterling terrace deposits, however, suggests an abatement of the Lake Bonneville outflow before the Cedar Butte Basalt dam was breached. If the boulder bars were formed after breaching of the lava dam they probably were emplaced by outflow waters of the Bonneville stage of spillover. If, however, they were emplaced during an earlier spill, then no record of spillover waters of the Bonneville stage has been recognized in this area.

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Reconnaissance Study of Quaternary Faults in and South of Yellowstone National Park, Wyoming

Abstract: Normal faults offset a bedrock surface scoured by Pleistocene ice in several areas within and south of Yellowstone National Park. Recurrent earthquake shocks and fresh appearance of some scarps suggest that movement is continuing along some faults. Four systems of faults are described.

Quaternary movement occurred along more than 60 faults on the Mirror Plateau, 15 miles northeast of Yellowstone Lake. Faults trend northwest, and several are more than 6 miles long. Maximum displacement exceeds 250 feet. The majority have northeast blocks downdropped, but some grabens and horsts are present. Eocene to Pliocene igneous or pyroclastic rocks are displaced. Ice moved southwest and south from the Beartooth and Absaroka ranges, nearly at right angles to the fault trends. Drainage in many ice-scoured valleys was disrupted by faulting, and small lakes (such as Mirror Lake) formed on downthrown blocks. Thermal activity occurs along some of these faults.

Directly east of Mirror Plateau, the Lamar normal fault has a displacement of 1300 + feet; perhaps 1000 feet of this may have occurred during Quaternary time.

The Yellowstone Falls fault system cuts Pliocene rhyolite southeast of the Upper Falls of the Yellowstone River. Faults trend northwest; maximum displacement exceeds 200 feet.

The Solfatara fault system trends north-northwest, cuts Pliocene rhyolite, and has a maximum Quaternary displacement of about 200 feet. The Hering Lake fault system is a northern extension of the Teton fault, trends northward, and cuts Pliocene rhyolite and rhyolitic welded tuff. Maximum displacement is about 200 feet. West-flowing streams established on bedrock scoured by ice were disrupted, and Beula, Hering, and South Boundary lakes formed on the downthrown (east) blocks.

The sharp angular unstepped appearance of fault scarps 50 to 200 feet high in these fault systems suggests that each scarp of this type was formed by one continuous movement. The displacement along faults associated with the Hebgen earthquake of August 1959 is commonly less than 20 feet. The abundance of Quaternary faults and the record of 18 earthquakes in historic time suggest that additional faulting and earthquake activity can be expected in the future. Recognition of this probability should influence the location and type of construction of buildings and other facilities.

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INTRODUCTION

This paper summarizes data on four Quaternary fault systems and several separate faults in and south of Yellowstone National Park, Wyoming (Fig. 1). Most of these have not previously been described. They are located in densely wooded areas with rough terrane. During parts of 14 field seasons, reconnaissance was done by plane, helicopter, boat, horse, jeep, and on foot. Investigations were terminated before the Hebgen earthquake (Witkind, 1959; Matthews and Roberts, 1960) of August 17–18, 1959, in and west of Yellowstone National Park, and subsequently there was no opportunity to examine the fault areas for evidence of renewed movement.

As a result of the Hebgen earthquake, with its attendant loss of life, property damage, and conspicuous crustal movements, both scientific and popular attention has been drawn to studies of Quaternary faults. Those in and south of Yellowstone National Park provide data that help place the Hebgen earthquake and its associated phenomena in their proper geologic perspective.

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The writer is greatly indebted to John de la Montagne for assistance in this study and for providing a perpetual challenge to ruling theories. G. M. Richmond made available his unpublished maps of glacial debris in Jackson Hole and has given valuable counsel throughout the investigation. Mr. and Mrs. Eliot Davis of the National Park Service have facilitated the work during many field seasons, in both Grand Teton and Yellowstone national parks. W. B. Hall and J. M. Hill provided a plane and took many photographs of faults and associated glacial phenomena. The writer appreciates the opportunity arranged by Mr. Hall to review in the field the evidence for high-level glaciation in the Gallatin Range. J. M. Lawson, William Kennedy, and Arthur Gervais of the U. S. Geological Survey made available a helicopter for use in less accessible areas of Yellowstone National Park during the summer of 1958. E. B. Leopold made pollen determinations from samples of post-rhyolite lacustrine strata in the Yellowstone Falls area.

The studies were conducted and the report prepared with the cooperation of the Geological Survey of Wyoming, the Department of Geology, University of Wyoming, and the U. S. National Park Service.

PREVIOUS WORK

The literature dealing with Quaternary faults and modern earthquakes in and south of Yellowstone National Park is meager, whereas that on glaciation is much more extensive. Howard (1937) described post-rhyolite faulting in several areas of Yellowstone National Park but did not consider it to be postglacial. Pardee (1950, p. 369–379) discussed postglacial faulting in the Yellowstone Valley north of the Park boundary, and along the east side of the Madison Valley, 15 miles west of the Park. Several of the Quaternary fault systems within the Park have been briefly mentioned (Love, 1959, p. 1782).

In Jackson Hole, there are two major normal faults of late Cenozoic age, the Teton fault on the west side of the Hole and the Hoback fault on the southeast. Both show some post-glacial movement (Love and Montagne, 1956, p. 169–178). A northern extension of the Teton fault continues a short distance into Yellowstone National Park (Fig. 1). Blackwelder (1915, p. 214–215) considered the east face of the Teton Range to be a fault-line scarp formed not later than middle Tertiary time. Horberg (1938, p. 64) and Edmund (1951, p. 65) considered part of it to be Recent. Fryxell (1938) described postglacial faulting along the east flank of the Teton Range. Love and Montagne (1956) discussed evidence for post-glacial movement along these faults and for tilting of Jackson Hole during Recent time. Lavin and Bonini (1957) made a reconnaissance gravity investigation along the Teton fault.

Unpublished studies of late Cenozoic faults in the Madison Canyon area (Fig. 1, loc. A) and Mount Sheridan area (Fig. 1, loc. B) have been made by F. R. Boyd (1957, Ph.D. thesis, Harvard Univ.). C. W. Brown (1957, Ph.D. thesis, Princeton Univ.) made similar studies in the northern part of the Park (Fig. 1, loc. C).

SUMMARY OF GLACIATION

Some of the late Cenozoic normal faults south of Yellowstone National Park have displacements of 10,000 feet or more (Love, 1956b), and movements have continued to Recent time. Therefore, to distinguish Tertiary and Quaternary faults, it has been necessary to study the distribution of glacial ice. Figure 2 shows glaciated and unglaciated areas, directions of ice movement, and, in the southwestern part of Yellowstone National Park,

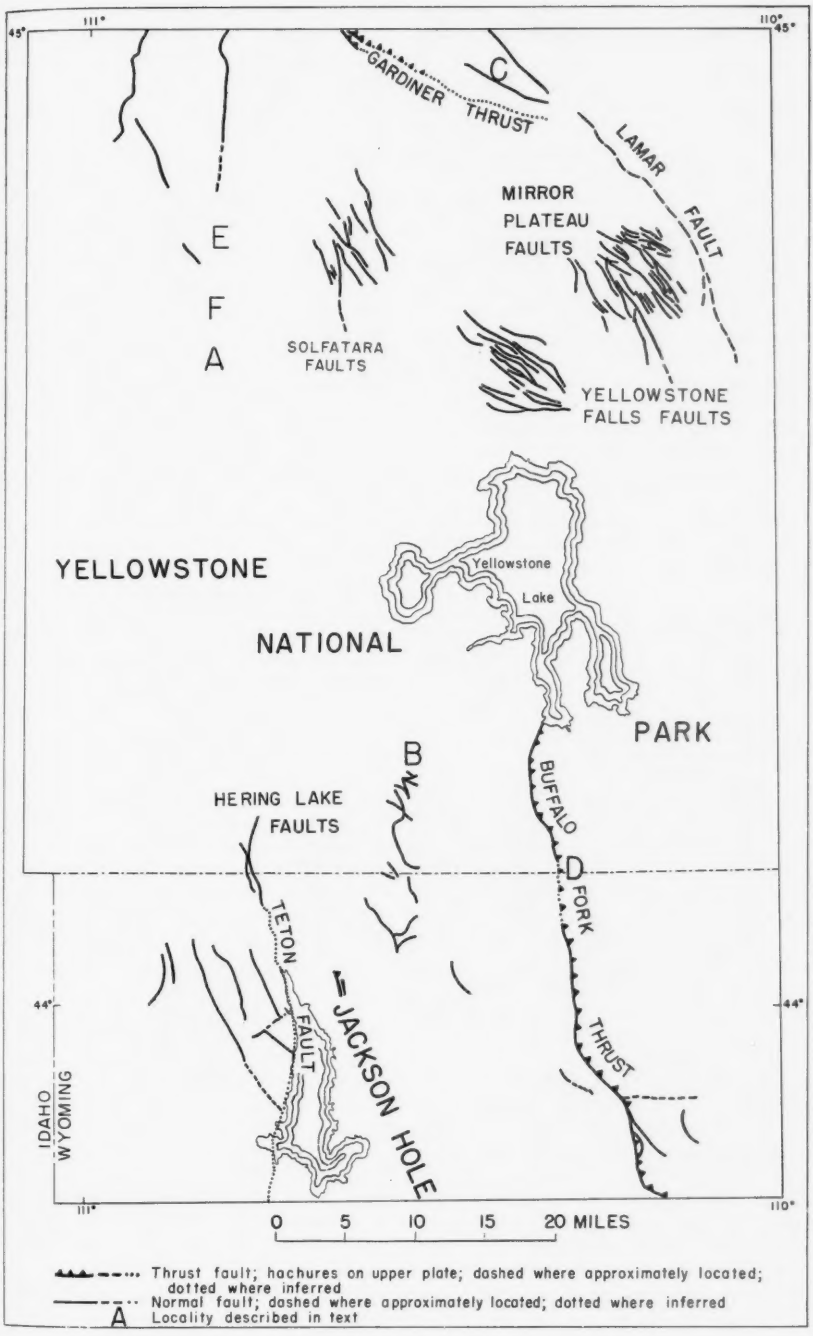

Figure 1. Faults in and south of Yellowstone National Park

two large areas of rhyolite extruded during an interglacial stage and not subsequently covered by ice.

Several times during the Pleistocene epoch, ice accumulated and moved westward, southwestward, or southward across the area shown in Plate 2. Inasmuch as the distribution pattern of the ice changed from one stage to another, the scoured bedrock surface is of different ages in different places. At many localities, this surface was subsequently offset by faults.

Directions of ice movements were determined by ice grooves, scour lines, and erratics of known source. A knowledge of direction of ice movement is in some places essential to recognition of Quaternary faults. For example, in areas of poor exposures, the steep side of an ice-cut valley is difficult to distinguish from a Quaternary fault scarp if they are parallel, but, if they are at right angles, fault offsets of the scoured bedrock surface are more readily apparent.

Holmes (1881) first recognized that a large ice mass once covered much of Yellowstone National Park. Later (1883, p. 55) he stated:

> "It takes no great effort of the imagination to picture the whole Park Plateau covered with a mass of snow and ice—a great *nevé* field fed by the accumulating snows from the surrounding highland, and extending its numerous ice tongues far out to the south and west, forcing them across the ocean divide into the Snake River Valley, or into the headwaters of the Missouri."

Weed (1893, p. 14) named the northern part of the ice mass the Yellowstone glacier. Hague (1896, p. 3) called the ice that moved southward into Jackson Hole the Snake glacier.

Howard (1937, p. 81–147) made a detailed study of glaciation in Yellowstone National Park, and the data supplied by him have been invaluable in all subsequent studies. He described and named the Lamar and Upper Yellowstone glaciers but did not discuss how they correlate with glacial stages in Jackson Hole.

Blackwelder (1915, p. 321–333) first named the Buffalo and successively younger Bull Lake and Pinedale glacial stages. The type locality of the Buffalo stage is along the eastern margin of Jackson Hole. This stage represents the most widespread accumulation of ice in the Jackson Hole area and is generally regarded as pre-Wisconsin. Older glaciation may have occurred but has not been definitely recognized.

The next younger stage is the Bull Lake, which is restricted to canyons and adjacent major valleys in Jackson Hole, but apparently was much more widespread in Yellowstone National Park. The extent of these deposits in the Park has not been determined. They may be of Wisconsin age. Carbon from a laminated silt apparently trapped behind a Bull Lake moraine in Jackson Hole has an age of about 27,000 years (Love, 1956b, p. 149). The moraine is certainly somewhat older.

The Pinedale glacial stage is the youngest in Jackson Hole and is restricted to canyons and valley floors adjacent to the high mountains. The ice of correlative age in Yellowstone National Park was more widespread, but its boundaries have not been determined.

Most of the glaciation above the 9000-foot contour shown on Figure 2 probably represents Blackwelder's Buffalo glacial stage, for that was the time of thickest and most widespread ice accumulation for which there is definite evidence in this region. During the younger stages, ice was thinner and less extensive, and in most places the maximum upper surface of accumulation was below 9000 feet.

The record of glaciation along the western margin of Yellowstone National Park has not been studied in detail. Hall (1959) recognized two and possibly three glacial stages in the Gallatin Range (Fig. 2, loc. G). During the earlier of these, ice was very thick and extensive and probably represents part of the Buffalo glacial stage.

Thick widespread rhyolite flows emerged from vents on the Madison Plateau (Fig. 2, loc. H) and spread westward beyond the west edge of Yellowstone National Park. The major flow mass overrides moraines thought to be part of the Bull Lake glacial stage (G. M. Richmond, written communication, 1959). The margins of the flow mass are, in places, modified by minor glaciation, probably of the Pinedale glacial stage.

The Pitchstone Plateau (Fig. 2, loc. I), in the southwestern part of Yellowstone National Park, is likewise composed of an interglacial rhyolite-flow sequence. Large glacial grooves, probably cut by ice of the Buffalo or Bull Lake glacial stages, extend from beneath the rhyolite on the southwest side of the Plateau, near the head of the east fork of Mountain Ash Creek (Fig. 2, loc. J). The margins of the Pitchstone flows have been modified in a few places by minor glaciation, probably of the Pinedale glacial stage.

Fryxell (1930, p. 1–125), Horberg (1938, p.

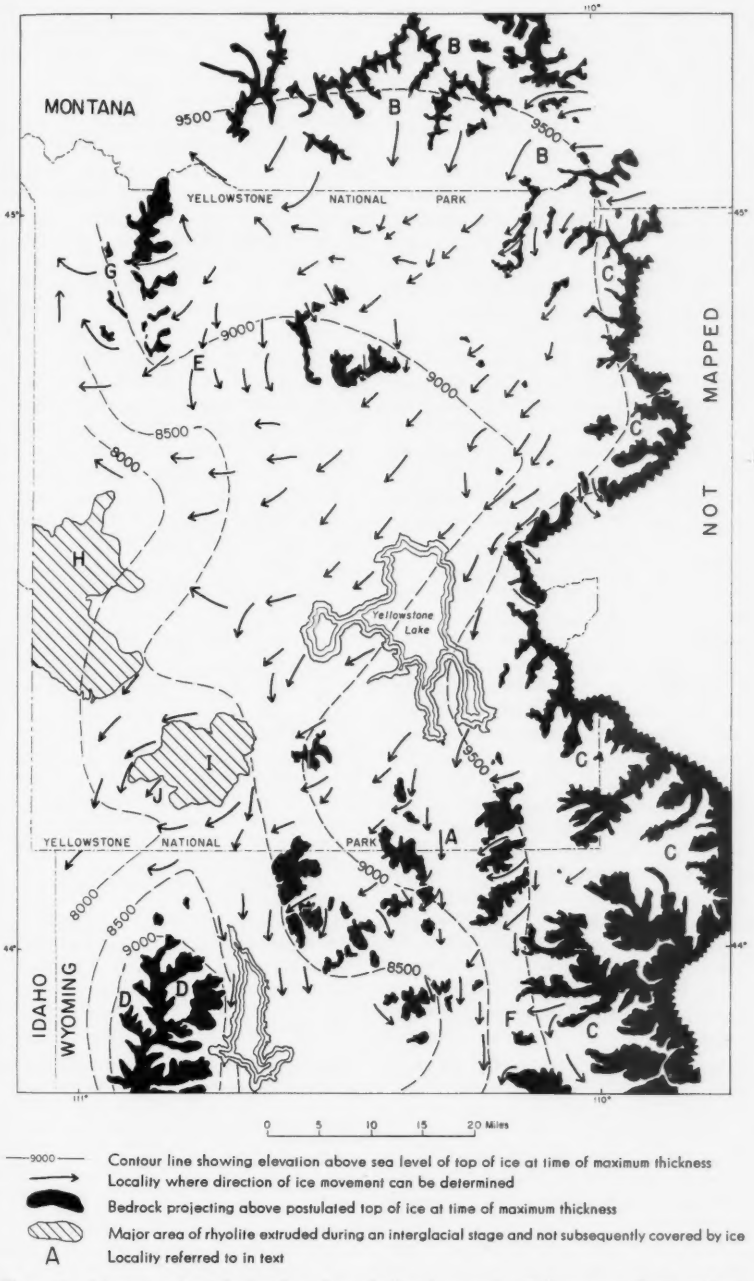

Figure 2. Map showing glaciated and unglaciated areas in Yellowstone Park region, direction of ice movement, and elevation above sea level of top of ice at time of maximum thickness

60–80), Edmund (1951, p. 67–78), Richmond (1948, p. 1400–1401), Love (1956b, p. 149–150), and Montagne and Love (1957, p. 1861) have described glaciation in Jackson Hole and adjacent areas.

Abundant large pink Precambrian granite erratics in and adjacent to south-trending glacial grooves and scour lines that cut bedrock of Cretaceous sedimentary rocks are found along the Yellowstone National Park boundary directly south of Yellowstone Lake (Fig. 2, loc. A). Three possible exposed source areas for these granite boulders are: the Teton Range, 25 miles to the southwest, Buffalo Fork Mountain, 18 miles southeast (Fig. 2, loc. F). and the Beartooth Mountains (Fig. 2, locs. B) 55 miles to the north. A fourth possible source area could be concealed under Yellowstone Lake, where Precambrian rocks might have been brought up along the northern part of the Buffalo Fork thrust fault (Fig. 1). Rocks as old as Ordovician are exposed along this fault (Love and others, 1955). A regional study of directions of ice movement (Fig. 2) indicates that the boulders probably came from the north, either from the Beartooth Mountains or from under Yellowstone Lake.

An analysis of the data indicates that within this area the major sources of ice accumulation were the Beartooth Mountains along the northern margin of Yellowstone National Park (Fig. 2, locs. B), the Absaroka Mountains along the eastern margin (Fig. 2, locs. C), and the Teton Range (Fig. 2, locs. D).

Ice during each stage cut a surface across hard and soft rocks ranging in age from Precambrian to Pleistocene. The final composite scoured surface has mountainous relief and in several areas, particularly those where Buffalo ice was thickest, is characterized by giant ice-cut grooves 1 to 5 miles long, 100 to 1000 feet wide, and 50 to several hundred feet deep. In many places in Jackson Hole slightly tilted Pleistocene lacustrine and fluviatile strata were overridden by ice.

QUATERNARY FAULTS

General Discussion

A basic assumption in determining the occurrence and magnitude of Quaternary faults is that any preglacial fault scarps trending at or nearly at right angles to direction of ice movement would have been obliterated or conspicuously modified by ice if they occurred below the upper limit of ice (Fig. 2). Therefore, in glaciated areas, unmodified scarps at right angles to direction of ice movement are considered to be fault scarps of Quaternary age.

Criteria used for recognition of Quaternary faults are: (1) linear offset of bedrock, (2) sharp, unmodified linear breaks in topography at or nearly at right angles to glacial scour lines, glacial grooves, or postglacial drainage; (3) lateral offsets of postglacial streams, (4) ponding, waterfalls, or rapids in postglacial streams; (5) distribution of hot springs coinciding with linear breaks in topography.

Almost all the Quaternary faults are in rhyolite or rhoylitic welded tuff, which characteristically supports dense stands of timber (Pl. 1), and which weathers so readily that exposures are poor. The faults were mapped on air photos and topographic maps. Field studies were made wherever circumstances permitted. Specific features of these faults are rather difficult to observe on the ground, however, because at most localities scarps appear as steep slopes covered with shattered rock and abundant trees (Pls. 2, 3, 4).

A generalized summary of Quaternary chronology follows (younger to older):

Normal faulting; the episode described in this paper. Faulting probably occurred during several intervals of Quaternary time, but glaciation has modified or obscured all but the youngest fault scarps in most places.
Minor sedimentation in local traps
Pinedale glaciation
Minor sedimentation in local traps
Extrusion of rhyolite flows constituting upper surfaces of Madison and Pitchstone plateaus
Minor sedimentation in local traps
Bull Lake glaciation
Minor sedimentation in local traps
Buffalo glaciation
White lacustrine and fluviatile strata, deposited in local downfaulted or downwarped basins
Complex interval of faulting, volcanism, erosion, deposition, and possibly glaciation; sequence of events and their dates are not adequately determined; they could be Pliocene or Pleistocene.

Major lithologic units of Tertiary age in and south of Yellowstone National Park are:

Pliocene
Rhyolite welded tuff, flows, and rhyolitic pyroclastic rocks; intertongues southward with middle Pliocene Teewinot Formation
Tower Creek Conglomerate; basalt flows
Miocene
Colter Formation; comprises a thick sequence of green and red mafic pyroclastic and igneous rocks 4 miles south of southwest margin of

Yellowstone National Park; may be equivalent to some basalt and andesite flows within the Park

Oligocene

"Late basic breccia" and underlying "late acid breccia," both of which appear to intertongue southeastward with Wiggins Formation

Eocene

"Early basalt flows"

"Early basic breccia"

"Early acid breccia"

Unnamed sequence of variegated claystone, sandstone, and conglomerate, northwestern Yellowstone National Park.

Paleocene

Pinyon Conglomerate.

Mirror Plateau Fault System

The Mirror Plateau fault system (Figs. 1, 3; Pls. 1, 2) consists of 60 or more northwest-trending normal faults, none of which has previously been mapped. Rocks range in age from late Eocene to Pliocene or possibly younger.

The "early basic breccia" (Iddings, 1896, p. 6; Hague, 1904) is the oldest pyroclastic sequence involved in faulting in this area. These rocks consist of volcanic breccia and conglomerate, composed of dark-green and brown pyroxene andesite and basalt fragments, interbedded with gray and green tuffaceous strata containing petrified trees and fossil leaves that are generally considered to be of late Eocene age. These rocks crop out along the northern and eastern margins of the area shown in Figure 3.

Overlying the "early basic breccia" are the "early basalt flows," which consist of dark-green to black dense hard basalt. At Locality A of Figure 3 (detail shown on Pl. 3 Fig. 1), the basalt is on the upthrown block and forms an escarpment about 100 feet high. A younger conglomerate, possibly the "late acid breccia," is on the downthrown block. In thin section the basalt is porphyritic and consists chiefly of plagioclase with lesser amounts of pyroxene, red-brown biotite, and magnetite. Hague (1904) mapped broad outcrops of the "early basalt flows" in the northern and eastern parts of the Mirror Plateau, and in the southeastern part of the area shown in Figure 3. The basalts are probably of late Eocene age, on the basis of comparison with similar basalt flows in the Tepee Trail Formation of late Eocene age in the Absaroka Range to the southeast.

Iddings (1896, p. 6) described the "late acid breccia" overlying the "early basalt flows," as consisting of light-colored hornblende-mica andesites with some fragments of mafic andesite. Hague (1904) mapped "late acid breccia" southeast of Mirror Lake and along the southern border of the area included in Figure 3. This sequence may be correlative with part of a similar sequence containing Oligocene fossils at Locality D of Figure 1 (Love, 1956a, p. 88).

Rhyolite flows and welded tuffs unconformably overlie all older rocks and consist of red and brown pumiceous to glassy rock with abundant vesicles, fragments of obsidian, and phenocrysts of quartz and plagioclase (Iddings, 1899, p. 391). On a fault scarp directly east of Mirror Lake (Fig. 3), the rhyolite is fine-grained, dense, porphyritic, and shows flow structure. Similar-appearing rhyolites and welded tuffs in the southern part of Yellowstone National Park extend southward into Jackson Hole and intertongue with the Teewinot Formation of middle Pliocene age (Love, 1956c, p. 1909; Houston, 1956, p. 133–139).

Either interbedded with or lying on the rhyolites in several places are light-colored tuffaceous deposits of unknown age, probably either Pliocene or Pleistocene. North of Wapiti Lake (Fig. 3) are extensive white, hot-spring deposits. In most places the rhyolite supports a dense growth of trees (Pl. 1), whereas both older and younger rocks support sparse trees.

Most of the faults cut only rhyolite flows and rhyolitic welded tuff in the area shown on Figure 3. The magnitude of displacement on the ice-scoured surface ranges from less than 20 feet to more than 350 feet. In general, the faults in the western part of the area, especially northwest and about 2 miles east of Wapiti Lake, have the greatest amount of displacement (Pl. 2, figs. 1, 2). The northeast block is downdropped along about two-thirds of the faults, but there are also many conspicuous grabens and horsts. Inclination of fault planes is rarely determinable but is probably steep.

Ice moved southwestward across broad saddles in the Mirror Plateau, and many southwestward-flowing tributaries to Broad Creek and Deep Creek were established along ice-cut valleys. Quaternary faults cut nearly at right angles across these valleys and, depending on the direction of fault displacement, have ponded, offset, or oversteepened the stream channels. Mirror Lake, 1400 feet along, and Wapiti Lake, 2000 feet long, were formed in this manner (Fig. 3).

Hot springs emerge along many fault planes. They are most abundant and conspicuous in the area northwest, north, and northeast of Wapiti

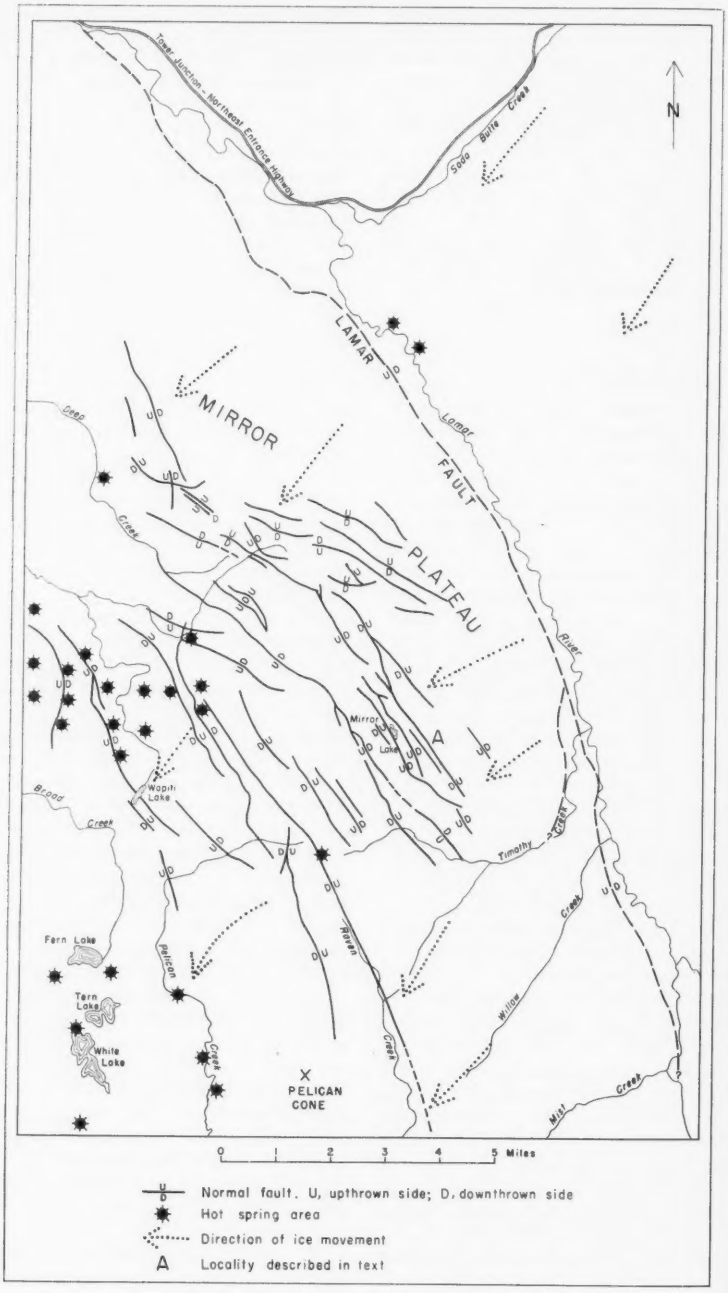

Figure 3. Lamar and Mirror Plateau faults

Lake (Pl. 2, fig. 2). Springs are most numerous in the general area of greatest fault displacement.

Most of the fault scarps appear to be fresh and undissected, although a few have rounded crests. Even some of the freshest support fairly large trees. A fresh scarp is moderately well exposed about a mile east-southeast of Mirror Lake (Fig. 3, loc. A; Pl. 3, fig. 1). Displacement of the ice-scoured surface is at least 100 feet,

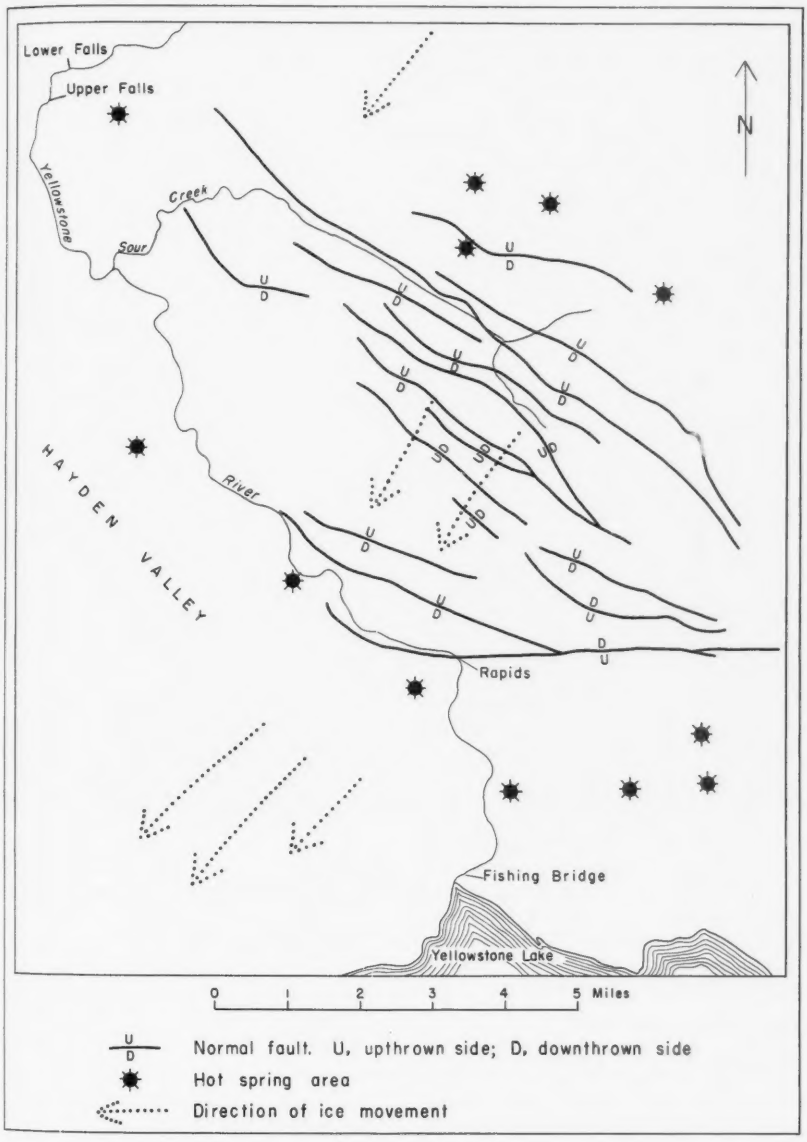

Figure 4. Yellowstone Falls fault system

and total displacement on the fault may be considerably more. Basalt is on the upthrown block. The downthrown block comprises horizontal beds of relatively fine-grained volcanic conglomerate of uncertain age and correlation, possibly the "late acid breccia." The conglomerate is composed of angular to rounded fragments of gray to reddish porphyritic andesite. In the one thin section studied, plagioclase and red-brown biotite are the only phenocrysts.

Inasmuch as the configuration of scarps ranges from fresh and angular to rounded, it is logical to assume that in the Mirror Plateau fault zone recurrent movement has gone on, perhaps from preglacial post-Pliocene rhyolite time to Recent. The faults shown on Figure 3 are those along which there was displacement during Quaternary time.

Lamar Normal Fault

The Lamar normal fault (Fig 3; Pl.1), although poorly exposed, is one of the longest (more than 20 miles) and may have one of the largest displacements (1300 feet or more) of the normal faults in Yellowstone National Park. Howard (1937, p. 71–74) inferred its presence partly on the basis of displaced or missing rock sequences and partly on physiographic evidence. Approximately the same stratigraphic units are present as are on the Mirror Plateau, except for some post-rhyolite basalt that crops out along the floor of the Lamar valley. The Pliocene rhyolite and rhyolitic welded tuff on the northeast block appear to have been down dropped as much as 1300 feet with respect to the southwest (Mirror Plateau) block. The age of the major movement is not known. Howard comments on the freshness of the scarp; however, some broad valleys, interpreted by the writer as having been cut by ice, extend from the upper part of the scarp toward the southwest across the Mirror Plateau. Some of the movement along the Lamar fault probably was late Tertiary, but if the differential elevation of ice-scoured surfaces cut on the rhyolite on the upthrown and downthrown blocks is significant, there may have been as much as 1000 feet of movement during the Quaternary.

Yellowstone Falls Fault System

The Yellowstone Falls fault system (Fig. 4) is about 7 miles southwest of the Mirror Plateau fault system and has not previously been described or mapped. The faults are longer (as much as 9 miles), fewer in number (16), have more displacement, and their scarps are more rounded than those on the Mirror Plateau. The Yellowstone Falls faults lie within an area 6 miles wide and 12 miles long, extending southeastward from a point about 2 miles southeast of the Upper Falls of the Yellowstone River. The faults are normal, appear to have nearly vertical planes, and the northeast blocks along about half the faults are downdropped.

The faults cut rhyolite and welded tuff of Pliocene or early Quaternary interglacial age. The age assignment of the rhyolite is based on the following pollen and spores, collected from the fine-grained diatomite, claystone, and tuff overlying the main mass of rhyolite on the east side of the canyon between the Upper Falls and Lower Falls of the Yellowstone River (identified from seven sediment samples from U. S. Geological Survey Paleobot. loc. no. D1212, by Estella B. Leopold, written communication, 1960):

Trees:
- *Abies* cf. *A. lasiocarpa*
- *Abies* cf. *A. concolor*
- *Picea* cf. *P. engelmanni*
- *Tsuga* cf. *T. mertensiana*
- *T.* cf. *T. heterophylla*
- *Pinus* spp.
- cf. Taxodiaceae
- *Quercus*

Shrubs:
- *Betula*
- *Alnus*
- *Salix*
- Chenopodiaceae: *Sarcobatus vermiculatus*
  cf. *Chenopodium album*
- Compositae: *Artemisia* spp.
  cf. *Ambrosia*
  cf. *Chrysothamnus*
- *Ephedra* cf. *E. nevadensis*
- *E.* cf. *E. torreyana*

Herbs:
- Caryophyllaceae: *Stellaria*
- Polemoniaceae: *Polemonium*
- Polygonaceae: *Eriogonum*
- Onagraceae: *Oenothera*
  Epilobium
- *Typha latifolia*
- Cyperaceae cf. *Scirpus*
- Gramineae

Fern allies:
- *Lycopodium*, *L. complanatum* group
- *Selaginella* cf. *S. densa*

High-altitude vertical photo shows Lamar River (A), Mirror Plateau (B), Mirror Lake (C), Deep Creek (D), and Quaternary fault scarps that have dammed or offset drainage. Details of faults are shown on Figure 3. Width of photo is approximately 8 miles; north is at top. Photo by Army Map Service.

PART OF MIRROR PLATEAU FAULT SYSTEM

Figure 1. Graben north-northeast of Wapiti Lake. View looking south-southeast at ponded and deflected drainage. Rounded scarps are in Pliocene rhyolite and are about 350 feet high.

Figure 2. Hot springs emerging along fault scarp (A) and other large hot springs (B) in graben. View looking north along scarp 1 mile north-northwest of Wapiti Lake. Height of scarp is about 200 feet.

NORMAL FAULTS IN THE WESTERN PART OF THE MIRROR PLATEAU FAULT SYSTEM

Figure 1. Fault scarp southeast of Mirror Lake. Displacement is about 100 feet. Upthrown block is composed of "early basalt flows", and downfaulted valley floor is of conglomerate, possibly "late acid breccia."

Figure 2. LeHardy's Rapids in Yellowstone River along trace of fault in Yellowstone Falls fault system. Horizontal rhyolite is exposed in river bed and on far bank above rapids.

FAULT SCARPS IN THE MIRROR PLATEAU AND YELLOWSTONE FALLS FAULT SYSTEMS

Figure 1. Air oblique photo showing Hering Lake faults. View looking northwest shows straight-line scarp along west side of Hering Lake (center) and a second scarp farther west in trees. Photo by W. B. Hall and J. M. Hill

Figure 2. West shore of Hering Lake showing cliffs and wooded slope along fault scarp of rhyolite that rises 150 feet above lake

FAULT SCARPS IN THE HERING LAKE FAULT SYSTEM

Algae:
Botryococcus
Pediastrum
diatoms, undet.[1]

Pollen of sage and pine predominate in these collections.

Displacement along several faults in the Yellowstone Falls fault system is as much as 300 feet, and on one fault there apparently was more than 600 feet of movement, some of which preceded the most recent glaciation. The last ice mass to occupy this area moved southwestward and cut giant grooves across rhyolite fault blocks that were beginning to rise. After the ice melted, the southwestward-flowing tributaries to the Yellowstone River flowed in these grooves. Renewed faulting, however, prevented most of these tributary streams from maintaining their courses across the rising fault blocks and caused diversion to the northwest along northwest-trending fault valleys. Thus, a rectilinear pattern was developed, with abandoned giant ice-cut grooves and stream-cut valleys extending at right angles across rounded glaciated fault-scarp ridges of rhyolite. Some of the grooves and abandoned stream channels are displaced vertically 200 feet or more, and this is interpreted as a rough measure of the magnitude of fault movement after the last ice melted in this area.

One fault crosses the Yellowstone River 3 miles north of Yellowstone Lake (Fig. 4), where the river is diverted westward, at right angles to its upstream course. Rapids are present along the fault trace (Pl. 3, fig. 2), and the rhyolite exposed on the downthrown side in the river bank is highly fractured. Some hot springs are present in the vicinity of the faults but are not so closely coincident with the fault traces as in the Mirror Plateau area.

White lacustrine and fluviatile strata, which characteristically support little vegetation in Hayden Valley, (Fig. 1; also Howard, 1937, p. 92–100) do not extend north of the Yellowstone Falls faults. The relationship of these strata to the faults has not been determined. Howard concluded (1937, p. 98) that the white sediments are younger than the glacial ice and were deposited in a lake (Howard's Hayden Lake) blocked from flowing north by remnants of the major ice mass. The lake in which the white strata were deposited may also have been formed in part by downwarp of Hayden Valley and in part by rising fault blocks in the Yellowstone Falls fault system that blocked or inhibited drainage to the north.

Solfatara Fault System

The Solfatara fault system (Fig. 5) is about 8 miles northwest of the Upper Falls of the Yellowstone River and has not previously been described or mapped. The present study only delimited a few of the most important faults; many others are present. They range in length from 2 to 10 miles, trend northwestward, and in most the northeast blocks are downdropped. Displacements commonly range from 40 to 120 feet, with a maximum of 200 feet. All the faults cut Pliocene rhyolite except the easternmost, which cuts some "early basic breccia." Extensively decomposed obsidian welded tuff crops out on the Solfatara Plateau. Considerable thermal activity is present in the vicinity of the Solfatara faults, but vents have not been mapped in sufficient detail to determine if the activity is related to the faults.

Ice moved southwestward across most of the Solfatara Plateau, but coalescing and crossing of undrained channelways suggest that ice masses from the north and east were competing for position along the western and southern margins of the area. The dominant postglacial drainage was southwestward, and this has been disrupted by the northwest-trending faults. Ice Lake, Wolf Lake, and Grebe Lake (Fig. 5) probably were formed by ponding of streams against rising fault scarps.

Hering Lake Fault System

The Hering Lake fault system (Figs. 1, 6; Pl. 4) is at the north end of the Teton normal fault. The most clear-cut evidence of faulting is along the west side of Hering Lake. The southern part of this system is shown on the Teton County geologic map (Love, 1956d); other than this, no published maps show these faults. The faults are confined to rhyolitic welded tuff which intertongues with white pumicite and tuff of the middle Pliocene Teewinot Formation. Displacement on the middle part of each of the main faults is about 150–200 feet (Fig. 6), with the east blocks downdropped.

Ice moved southwestward across this area,

[1] After this paper was submitted for publication, a report was received from K. E. Lohman, listing 35 genera and species of diatoms identified from these samples. He considers the diatoms to be of Pleistocene age.

as is shown by scour lines and erratics of lithologic types known only from localities to the east-northeast. Falls River and its tributaries were established on this surface and were subsequently ponded or offset by fault blocks that rose across their channels. Beula, Hering, and South Boundary lakes were formed in this way (Fig. 6; Pl. 4). The fault scarps are straight, vertically angular, and unmodified; they are covered with a heavy growth of timber.

Along the eastern margin of the area shown in Figure 6, Polecat Creek divides, with one part flowing south and another part seeping eastward into Crawfish Creek; these relationships suggest Recent faulting. However, in that locality any Recent fault scarp would probably

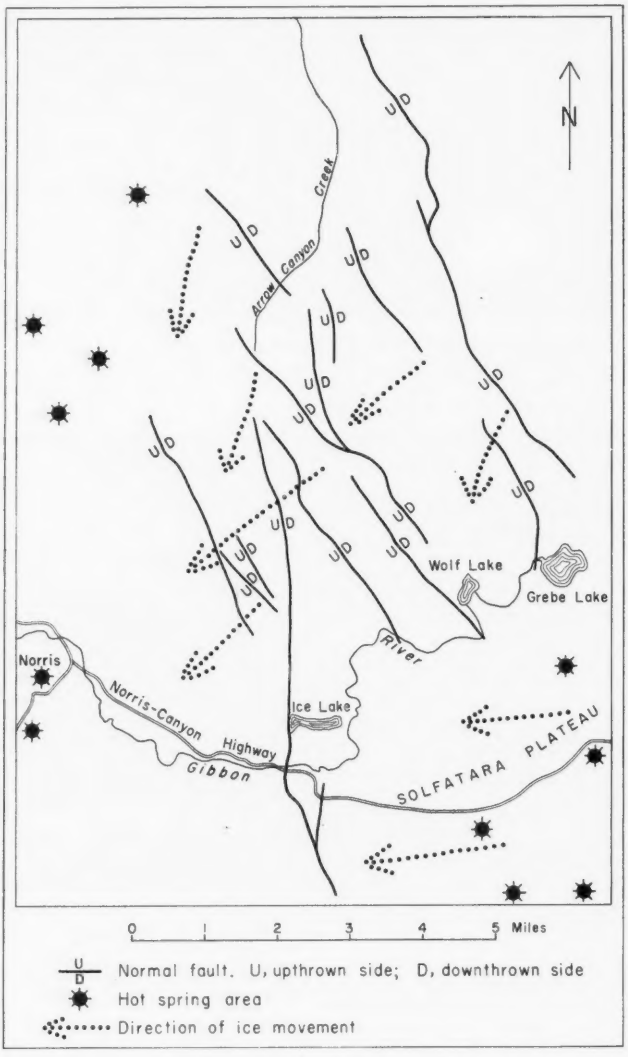

Figure 5. Solfatara fault system

parallel a glacial scour face cut in rhyolite and rhyolitic welded tuff along the southeastern margin of the Pitchstone Plateau. Such an orientation, coupled with relatively poor exposures, makes it difficult to distinguish a fault scarp from an unfaulted ice-scoured scarp. At the base of a scarp directly west of Polecat Creek (Fig. 6) is a row of 15 hot springs whose positions may be governed by a fault.

After the Hebgen earthquake of August 17–18, 1959, Crawfish Creek and Polecat Creek were turbid with yellowish-white mud. Polecat

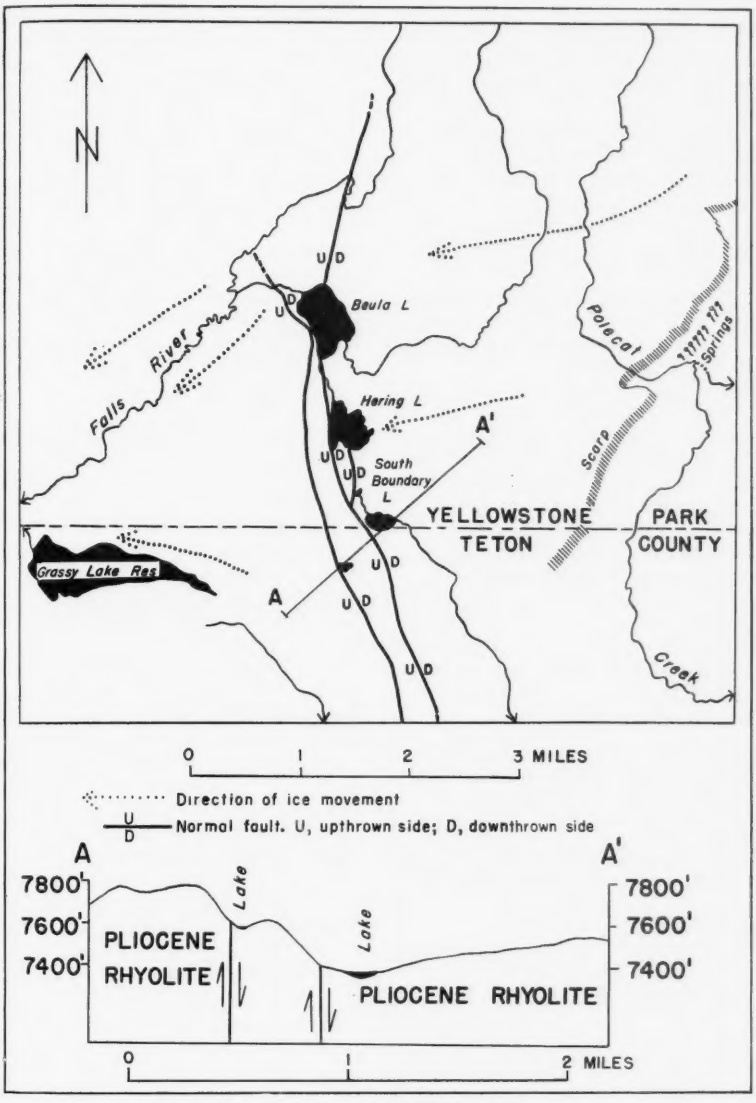

Figure 6. Hering Lake fault system

Creek continued to be turbid for several weeks, and the mud discolored the northern half of Jackson Lake, into which Polecat Creek empties. Several geologists who flew over the hot-spring area along the scarp reported greatly increased thermal activity. There was no opportunity to search for new scarps on the ground.

Probable Faults in Inadequately Studied Areas

Air photos show a number of areas of probable faulting that were not examined on the ground or were inadequately studied. One area (Fig. 1, loc. E) is a northwest-trending zone beginning about 2 miles southeast of Mount Holmes and extending southeast for about 5 miles to a point 4 miles northwest of Norris. Rocks in this area are chiefly upper Cenozoic rhyolite. Northeast blocks are downdropped along rounded scarps. Prominent glaciated valleys trend south-southwest across this zone (Fig. 2, loc. E). The valleys contain numerous erratics of Precambrian granite that probably came from the Beartooth Mountains. Precambrian rocks are exposed in the Gallatin Range to the northwest, but any ice bringing erratics from that direction would have moved nearly at right angles to the valleys and would certainly have modified them. Faults have not offset these valleys extensively, so most of the movement along the scarps apparently preceded the advance of this ice mass.

Another area of probable faulting (Fig. 1, loc. F) is about 2 miles west of the Gibbon Geyser Basin in dominantly upper Cenozoic rhyolite. Lineations on areal photographs trend north and north-northwest across ice-scour lines. The amount of postglacial movement appears to be small.

COMPARISON OF LOCAL EARTHQUAKE EFFECTS WITH THOSE FROM MAJOR HISTORIC EARTHQUAKES

Heck and Eppley (1958, p. 44–57) listed 9 earthquakes in northwestern Wyoming between 1856 and 1956. Woollard (1958, p. 1138) described 11 in the same area prior to 1938 and 7 more between 1938 and 1957. Some were multiple shocks, and one consisted of a long series of shocks (August 24 to December 22, 1930) (Heck and Eppley, 1958, p. 54). Two earthquakes of moderate intensity are listed as having occurred in Yellowstone National Park prior to the Hebgen earthquake of August 17–18, 1959. Blackwelder (1926, p. 196), Fryxell (1933), and Gale (1940) described earthquake shocks in Jackson Hole.

Byerly (1942, p. 80–102) and Davison (1936) discussed the great earthquakes of the world. The Indian earthquake of 1897 resulted in a maximum vertical crustal displacement of 35 feet, and the Mino-Owani earthquake of 1891 in Japan in a displacement of 20 feet. During the Kwanto earthquake of 1923 in Japan a maximum sinking of one part of the adjacent Sagami Bay was 690 feet, and an uplift of another part was 820 feet; it is not known how much of this movement was due to shifting of loose material and how much to tectonic deformation (Byerly, 1942, p. 101). During an earthquake in 1692, the seaward margin of Port Royal, Jamaica, subsided 30 feet (Link, 1960, p. 166–167). Most of the faults developed at the time of the Hebgen earthquake have less than 20 feet of displacement.

CONCLUSIONS

Faults along which there has been 100–300 feet or more displacement in Quaternary time are present in several parts of Yellowstone National Park. There is no way to determine how much displacement occurred in a single episode, but the sharp angular unstepped appearance of several scarps 50–150 feet high (Pls. 1, 4) suggests that each was formed by one major uninterrupted movement. This amount of movement is comparable or greater than that occurring during some of the major historic earthquakes of the world, and considerably greater than that along faults associated with the Hebgen earthquake.

The past history of faulting and earthquake activity in the Yellowstone National Park region, suggests that similar activity of significant magnitude can be expected in the future. Taking proper cognizance of this, tourist facilities and permanent installations can be located in areas safe from rock slides and constructed on bedrock, with adequate structural designs that will insure a minimum of earthquake damage and loss of life.

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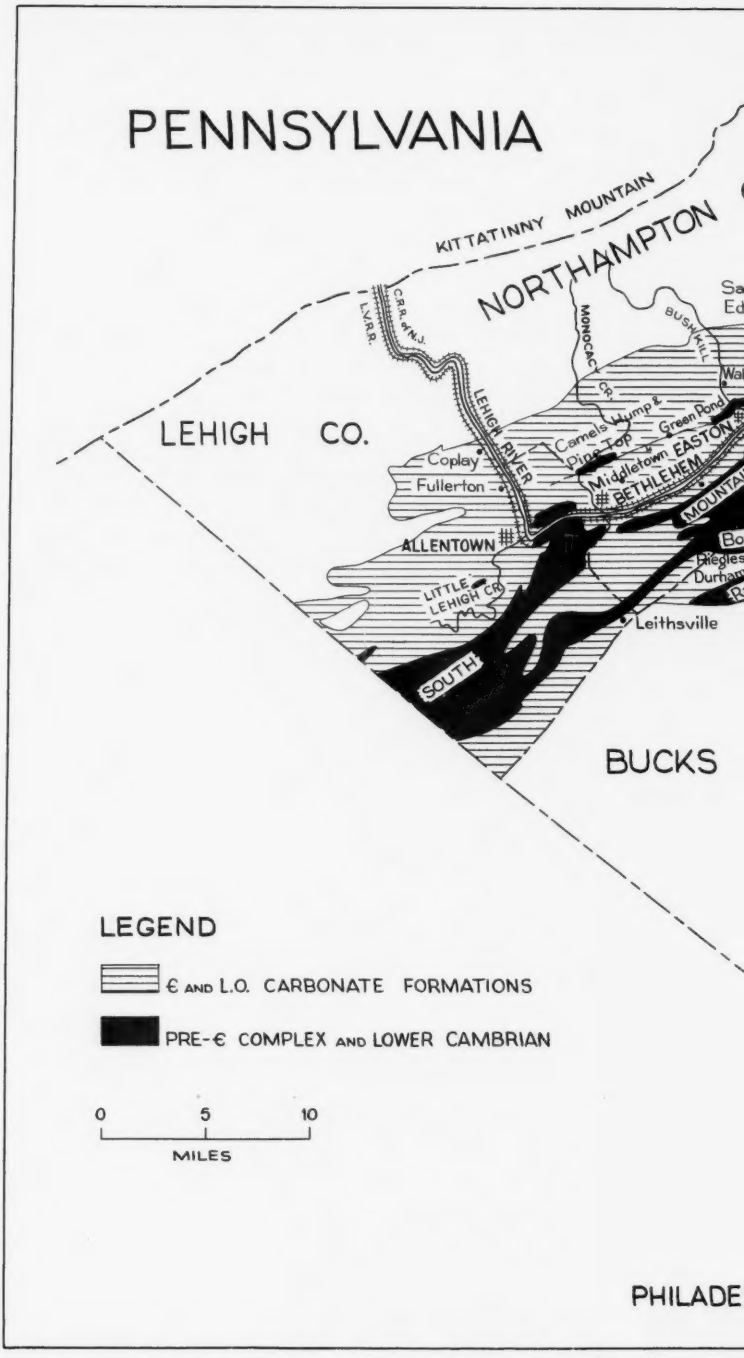
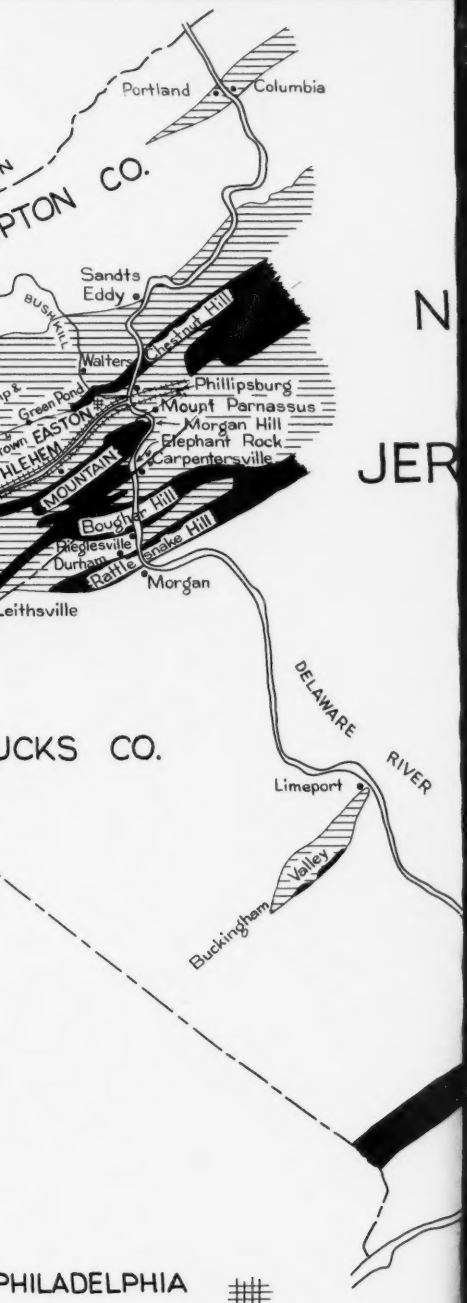
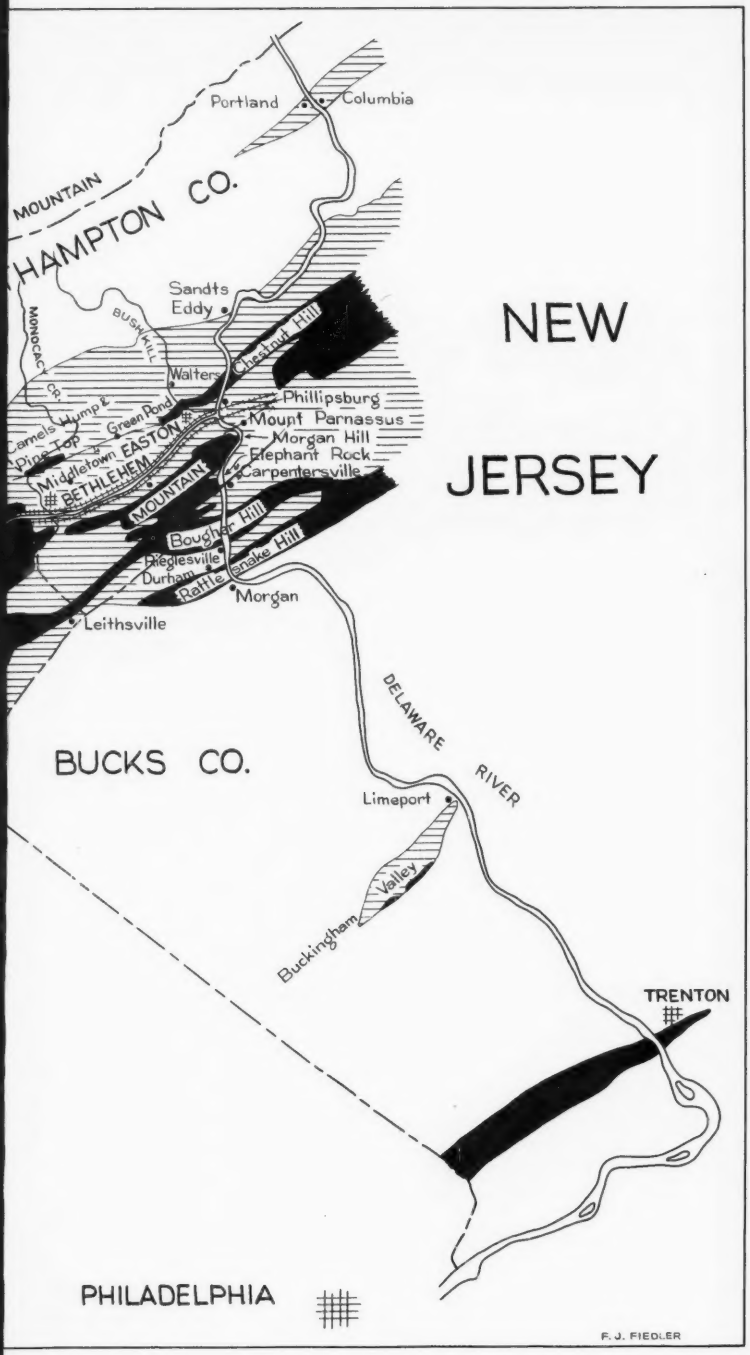


SKETCH MAP OF EASTERN PENNSYLVANI

WILLARD, PLATE 1
Geological Society of America Bulletin, volume 72



LVANIA AND ADJACENT NEW



NNSYLVANIA AND ADJACENT NEW JERSEY

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Stratigraphy of the Cambrian Sedimentary Rocks of Eastern Pennsylvania

Abstract: This paper describes the stratigraphy of the Cambrian sedimentary rocks of eastern Pennsylvania and adjacent New Jersey in the neighborhood of the confluence of the Delaware and Lehigh rivers. The succession is:

Upper Cambrian . . .	Allentown Formation Limeport Formation
Middle(?) Cambrian. .	Leithsville Formation
Lower Cambrian . . .	Hardyston Formation.

The sequence, except for the Hardyston, is predominantly carbonate rocks. The total thickness is about 3000 feet. The basal formation, the Hardyston, is clastic and changes abruptly lithologically and in thickness and was probably laid down in a rapidly advancing sea. The carbonate formations which follow, except the Leithsville, are fossiliferous. Partly through analogy with recent conditions, it is postulated that these carbonate rocks and their "cryptozoöns were deposited in a shallow, agitated sea into which little detritus was carried.

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INTRODUCTION

PROBLEM: This study deals with the Cambrian sedimentary formations in eastern Pennsylvania and adjacent parts of New Jersey. New data include the paleoecology of the stromatolites of the Upper Cambrian strata. Because the writer (1958) previously discussed the Lower Ordovician formations, only a summary paragraph is introduced.

PREVIOUS WORK: Wherry (1909) subdivided the Lower Paleozoic rocks of the Lehigh Valley. Hills (1935) recorded insoluble residues. Hills used, but B. L. Miller and coworkers (1934, 1939; 1941) partly rejected, Wherry's formational names. Howell, Roberts, and Willard (1950) revised the Cambrian nomenclature, reverted to Wherry's designations, and added one formational name. They applied Howell's study of invertebrates and Roberts's knowledge of the Upper Cambrian stromatolites. Willard (1955a) described Cambrian contacts. Willard (*in* Willard *et al.*, 1959) reported on the Cambrian rocks in Bucks County.[1] The present account deals chiefly with Northampton County, Pennsylvania, and refers to Lehigh County to the west, Bucks County to the

[1] The stratigraphic nomenclature in this book dealing with the Cambrian in Bucks County is not in agreement with that used here. Certain modifications were inserted therein in compliance with usage by the Pennsylvania Geological Survey.

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south, and a slice of New Jersey adjacent to the Delaware River.

ACKNOWLEDGMENT: This study was made possible by a grant from the Penrose Bequest of The Geological Society of America.

PHYSIOGRAPHIC SETTING: The area embraces part of the Appalachian Valley and Ridge Province. Low, east-trending ridges of Precambrian crystalline rocks are separated by limestone-floored valleys, in the southern half of the area. North thereof is a broad, limestone valley. The Lehigh River joins the Delaware at Easton (Pl. 1). Most of the territory described lies within 15 miles of that confluence. Many of the places mentioned are on the Easton 7½-minute quadrangle of the U.S. Geological Survey.

STRATIGRAPHIC SEQUENCE

General Statement

The stratigraphic sequence of the Cambrian and Lower Ordovician formations of the Lehigh Valley is:

		Feet
Lower Ordovician	Coplay("Beekmantown") Formation	1500
Upper Cambrian	Allentown Formation	900
	Limeport Formation	900–950
Middle(?)Cambrian	Leithsville Formation	800–900
Lower Cambrian	Hardyston Formation	0–200±

The Middle Ordovician Jacksonburg limestone overlies the Coplay ("Beekmantown"); the Hardyston rests on Precambrian metamorphic rocks. Excepting the Hardyston, all the formations are carbonates. They are distinguished chiefly upon lithology, but the Limeport, Allentown, and Coplay have diagnostic fossils.

Hardyston Formation

NAME, LITHOLOGY, AND THICKNESS: Kümmel and Weller (1902) named the Hardyston. Although many have designated it "the Hardyston quartzite," the lithologic changes vertically and horizontally make this inappropriate. The commonest rocks are silicarenite and orthoquartzite. Both are locally more or less feldspathic. In most exposures the grains are well-sorted and subangular. Conglomerate lenses are numerous, and they are not confined to the base. Most of the phenoclasts are milky quartz, are less than half an inch in diameter, and tend to be spherical and well-rounded. Ellipsoidal pebbles were found at one place. Feldspar pebbles mingle with the quartz at some outcrops. The matrix of the conglomerates resembles the more feldspathic sandstone fractions. Sericitic and nonsericitic shales are common where the formation grades into the overlying Leithsville.

Brown, yellow, or buff jasper is locally common, chert rare. Quartz veins are many. One pegmatite intrusion has been reported (Fraser, 1939; Virgin, 1955). Iron ores were once extensively exploited from the Hardyston. Pinite schist (Willard and Virgin, in press) separates the Hardyston from the Precambrian locally. In some sections the lower beds of the Hardyston contain pinite. The Hardyston breaks down to brownish, impure sand and kaolinite. The overburden is sandy.

The Hardyston makes little topographic expression. Confined to the lower flanks of the ridges of Precambrian rocks, it is at many places talus-hidden, but at others supports an ill-defined bench. No complete outcrop section

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PLATE 2. LOWER AND MIDDLE (?) CAMBRIAN, EASTERN PENNSYLVANIA

Figure 1. Cambrian-Precambrian contact. Excavation on Lehigh University campus; top, opposite truck, is Hardyston; boss at lower left, Precambrian granite gneiss; intervening soil at stack of tools

Figure 2. Basal Hardyston, conglomerate of ellipsoidal pebbles on Little Lehigh Creek above Fish Hatchery, 2 miles south of Allentown

Figure 3. Leithsville Formation; heavy-bedded, dolomitic beds on U.S. Route 22, west of Easton

PLATE 3. UPPER CAMBRIAN OF EASTERN PENNSYLVANIA

Figure 1. Limeport Formation, near middle, in abandoned quarry west edge of Easton. Note alternating light- and dark-gray beds

Figure 2. Limeport Formation, lower part, in cut on U.S. Route 22 west of Easton. Note alternate light- and dark-gray beds. Many of the light-gray beds are stromatolite-bearing

Figure 3. Allentown Formation adjacent to U.S. Route 22 on the Bushkill at Easton. Note absence of light- and dark-gray bedding which is characteristic of Limeport Formation

Figure 4. Disconformity (at hammer) between Allentown Formation above and Limeport Formation below. North side of Lehigh River near Freemansburg

Figure 1

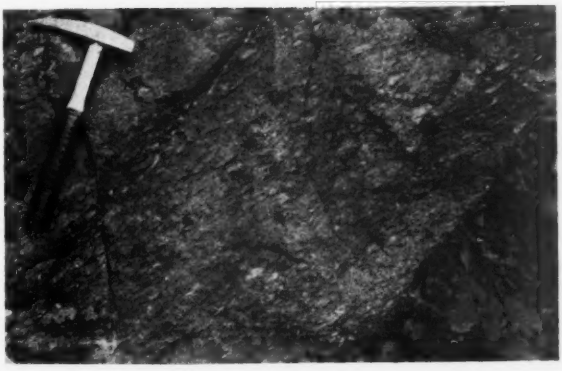

Figure 2

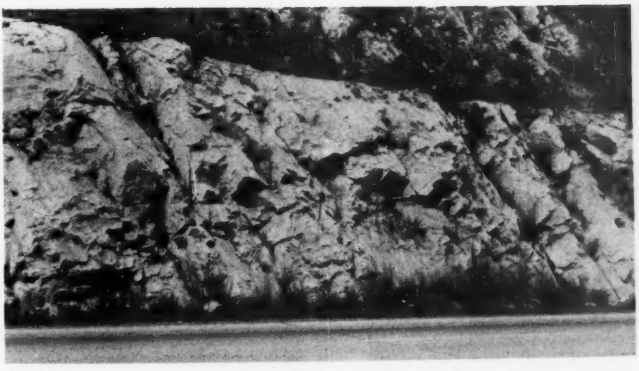

Figure 3

LOWER AND MIDDLE (?) CAMBRIAN, EASTERN PENNSYLVANIA

Figure 1

Figure 2

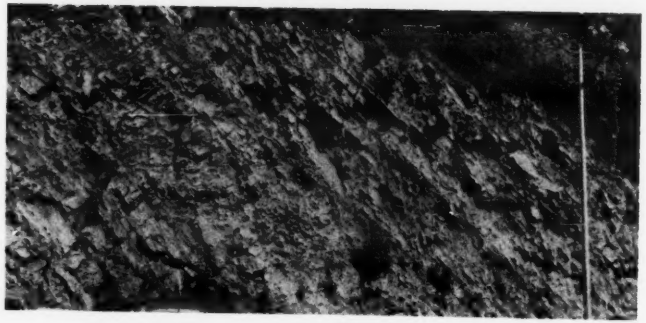

Figure 3

Figure 4

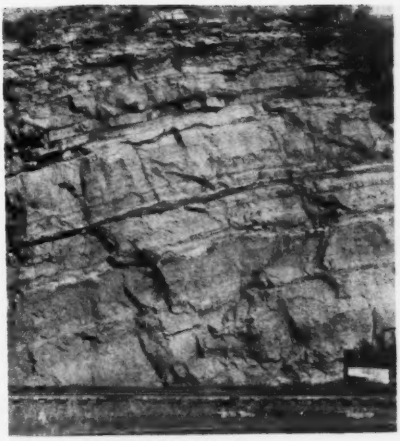

Figure 1

Figure 2

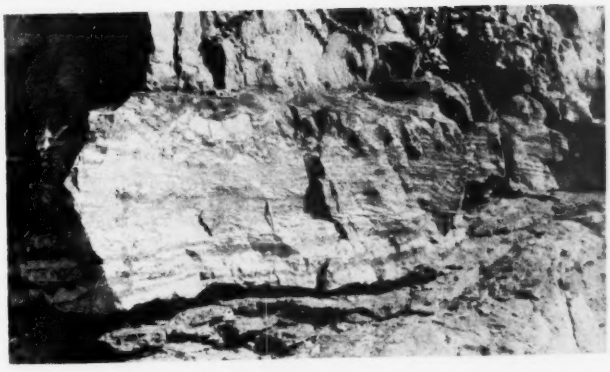

Figure 3

DETAILS OF LIMEPORT FORMATION NEAR EASTON, PENNSYLVANIA

Figure 2

Figure 4

Figure 1

Figure 3

has been seen in the region. The maximum thickness is about 200 feet, but locally the Hardyston is absent.

The writer (1955a) described the contacts of the Hardyston. Where pinite schist is absent at the base, the Hardyston may be welded to the Precambrian gneiss. Subsequent to the writer's description, excavations on the campus of Lehigh University exposed two places where the basal contact was visible (Pl. 2, fig. 1). At one cut an indurated residual, unstratified soil separated the gneiss (into which it graded) from the Hardyston. At the other, transported soil, also made up of derivatives from the gneiss, separated the Cambrian from the Precambrian. This second soil was bedded, sorted, and in sharp contact with the gneiss. Neither exposure carried pinite. The relief of the pre-Hardyston surface of the gneiss was about 13 feet.

FOSSILS: Organic remains are rare. Weller (1903) figured *Olenellus thompsoni* (Hall) from New Jersey. Wherry (1931) found fragments of it in Bucks County. In the Lehigh Valley the only reported fossil is *Scolithus linearis* (Haldeman) (Miller *et al.*, 1939; 1941).

CORRELATION: The equivalence of the Hardyston to the type in New Jersey is established (Miller *et al.*, 1939, p. 215). Howell *et al.* (1944) correlated it with the Chickies Formation of southern Pennsylvania and the Antietam Quartzite farther south and west.

LOCAL VARIATIONS: The writer has observed dense, dark-gray to bluish quartzite outcrops south of the Lehigh River east of Bethlehem, along the road west of Redington, on the Lehigh University campus at Bethlehem, and westward at the north foot of South Mountain. He found a conglomerate at the base at the Bucks-Northampton county line west of Springtown that is a mixture of quartz and feldspar pebbles up to 2 inches across. Above the Fish Hatchery on Little Lehigh Creek 2 miles south of Allentown, the basal 15–20 feet is conglomerate composed of 1-inch lavender quartz pebbles (Pl. 2, fig. 2). Along the east bank of the Bushkill at the water gap through Chestnut Hill west of Easton, conglomerate rests directly on gneiss. Here, the lower 2 to 3 feet consists of clear quartz pebbles up to 1 inch in diameter.

Leithsville Formation

NAME: The Leithsville is the poorest exposed and least understood of the carbonate Cambrian formations. No uninterrupted section has been discovered. The formation produces no noticeable topographic expression. The name Leithsville (pronounced "Litesville"), proposed by Wherry (1909), was used by Hills (1935), abandoned by Miller *et al.* (1939; 1941) in favor of "Tomstown", and reactivated by Howell, Roberts, and Willard (1950, p. 1357).

LITHOLOGY: The Leithsville is a high-magnesium carbonate rock. Fresh exposures are gray-, buff-, chalk-, or rust-colored. Some beds weather lighter, some darker, others turn yellow. Residual soil commonly contains chips of sericitic shale that was interbedded among the carbonate strata. Few caves have been discovered in the carbonate formations in the Lehigh Valley except the Leithsville, which is notably cavernous (Pl. 5, fig. 1). Several large caves are well known, and the writer found a number of small ones during field work. Besides the carbonate, the formation includes interbedded light-gray to brownish, sericitic shale and an occasional stringer of black chert.

PLATE 4. DETAILS OF LIMEPORT FORMATION NEAR EASTON, PENNSYLVANIA

Figure 1. Limeport Formation in railroad cut near confluence of Lehigh and Delaware rivers, Easton, illustrating ribbon bands

Figure 2. Details of ribbon band from Figure 2

Figure 3. Disrupted ribbon band in Limeport Formation east of Lehigh Valley Railroad Station at Easton

PLATE 5. CAVES AND STROMATOLITES, EASTERN PENNSYLVANIA

Figure 1. Cave openings, Leithsville Formation, at abandoned quarry, Redington, south of Lehigh River east from Freemansburg

Figure 2. Stromatolite bioherm, light-gray bed above dark-gray line, tops disrupted and pieces incorporated in overlying stratum; foot rule shows scale

Figure 3. Cast of tops of stromatolites on lower surface of bed along spur track north of Lehigh River and east of Freemansburg

Figure 4. Stromatolites smothered by shale which follows contour of heads; foot rule shows scale

Silicarenite is not common; quartz lenses and quartz-filled gash veins are numerous. Oölite is scarce.

CHEMICAL CONSTITUENTS: Miller (1934) published many chemical analyses of the Cambrian formations in eastern Pennsylvania. Recalculation of his analyses (1934, p. 235) in order to obtain calcite-dolomite ratios and application of Pettijohn's table (1949, p. 313), indicate dolomite.

$CaCO_3$	5.2	6.9	2.9	0.6
$CaMg(CO_3)_2$	90.7	85.8	90.5	94.8

INSOLUBLE RESIDUES: No insoluble-residue analyses of the Leithsville were run for this study. Hills (1935), from insoluble residue analyses, divided the formation into an upper and a lower part. He further subdivided the upper portion into three units:

(3) Shale, sand[2], and chert
(2) "Dark" shale, sand, and chert
(1) Many rounded quartz grains

From the lower part he listed "dark" shale, a little sand, and pieces of "dolomite".

BEDDING: Bedding varies from shale to strata 6-8 feet thick. Some larger exposures show vague rhythms of recurrent sequences of massive to platy to shaly bedding (Pl. 2, fig. 3)

FOSSILS AND PSEUDOFOSSILS: Mud cracks abound in the shaly beds, and ripple marks are numerous (Willard, 1954). Edgewise conglomerate is common, but it is seldom noticed in fresh rock. No identifiable fossils are known. Hills (1935) reported "Foraminifera". The writer infers that the Leithsville is Middle Cambrian. It is in transitional contact with overlying Limeport (early late Cambrian) and underlying Hardyston (early Cambrian), and it may include some beds of early and of late Cambrian ages.

THICKNESS: Thickness measurements are incomplete. Miller *et al.* (1939; 1941) assigned 900 to 1000 feet to the formation (his "Tomstown") in the eastern part of the Lehigh Valley. The writer calculated the thickness along the north side of the Bushkill west of Easton as between 800 and 900 feet.

Limeport Formation

NAME: Howell, Roberts, and Willard (1950, p. 1360-1361) named the Limeport Formation for Limeport, Bucks County, Pennsylvania. It is approximately synonymous with the lower half of Miller's Allentown Formation (Miller *et al.*, 1939; 1941). The Limeport was separated from the remaining Upper Cambrian on paleontological evidence and on the presence of a disconformity at its top (Pl. 3, fig. 4). Howell, Roberts, and Willard retained the name Allentown for the overlying Upper Cambrian beds.

LITHOLOGY: The Limeport is a dolomite. In most fresh exposures it is a uniform gray. In weathered outcrops and some fresh cuts, light- and dark-gray beds alternate (Pl. 3, figs. 1, 2). A few beds are buff, rarely pinkish. At the type locality in eastern Bucks County, the highest 40 to 50 feet are red shale and brown silicarenite interbedded with flaggy dolomite. Here, mud cracks and ripple marks are plentiful; white, oval rings and bands mark the stratal surfaces but are unrelated to the bedding (Willard *et al.*, 1959, p. 45).

The Limeport texturally may be dense, fine-grained, or crystalline. Oölite, locally pisolitic, is common throughout. Edgewise conglomerate or sedimentary breccia persist. Quartz grains, scattered or in thin lenses, are plentiful. Most particles are clear, glassy, well rounded, spherical, and well sorted. Some have frosted surfaces. The diameter of the grains rarely exceeds one-tenth of an inch.

Quartz-filled vugs are found, particularly near the top of the formation; calcite-filled vugs are rarer. Black-chert nodules show, but chert bands are few. Symmetric and unsymmetric ripple marks are plentiful throughout. Some have a wave length of as much as a foot (Willard, 1954). Mud cracks are confined to shalier beds. Some outcrops of the Limeport display a bluish stain from seeping water. Found rarely in the Allentown, this staining is presumptive evidence for the Limeport. The writer has also observed a white efflorescence on the beds but has not determined its chemical nature.

CHEMICAL CONSTITUENTS: The writer has had analyses by Miller (1934, p. 254) from the lower part of his "Allentown" recalculated to determine the calcite-dolomite ratios after Pettijohn's method (1949). The recalculations show the predominance of calcium-magnesium salt:

$CaCO_3$	2.3	18.3	5.5	36.0
$CaMg(CO_3)_2$	92.3	69.9	88.3	39.8

INSOLUBLE RESIDUES: Hills (1935), from his work on insoluble residues, divided the Allentown (old usage) much as did Howell, Roberts, and Willard in 1950. Hills's "lower Allentown"

[2] Presumably the "sand" reported is quartzose.

corresponds approximately to the Limeport. At its top the insoluble material was sand (presumably quartz), toward the middle shale fragments and another sandy zone were present, and near the base more sand and shale were found. R. H. Lessentine (1952, M. S. thesis, Lehigh Univ.) made insoluble-residue determinations on the Limeport in eastern

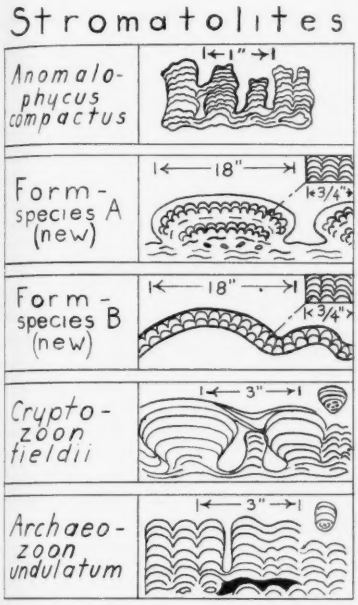

Figure 1. Key to the identification of Upper Cambrian stromatolites. Copied after Howell, Roberts, and Willard (1950)

Bucks County. He reported rounded and subrounded grains of quartz, mostly frosted, and also angular quartz grains and tabular shale fragments with angular bits of chert and feldspar. Lessentine did not mention any characteristic suites or zones of insoluble materials.

Kenneth B. Knowlton, a former graduate student at Lehigh University, at the author's request, ran insoluble-residue determinations on the topmost beds of the Limeport at Columbia, New Jersey, and reported rounded, fine quartz grains and clay particles. In the Nancy Run section at Middletown, Pennsylvania, the top of the Limeport yielded quartz grains and gray clay particles.

STRATIFICATION: The bedding of the Limeport is uniform. The alternating light- and dark-gray strata are 8–12 inches thick. Flaggy beds and shale are unusual. Here and there, weathered surfaces show laminations. A peculiar local feature of the Limeport is "ribbon limestone". Through 1–3 feet there appear alternating dark- and light-gray, almost white, beds 1–2 inches thick (Pl. 4, figs. 1, 2, 3). This variety is rare in the overlying Allentown.

FOSSILS: Howell collected Dresbachian invertebrates at the type locality, Limeport, Pennsylvania (Howell, Roberts, and Willard, 1950, p. 1360; Howell, 1957). Hills (1935) reported "Foraminifera", which, so far as the present writer knows, have not been described. Besides the invertebrates, stromatolites are plentiful (Fig. 1), including: *Cryptozoön proliferium fieldii* Fenton and Fenton, *Archaeozoön undulatum* (Bassler), and *Dolotophycus expansus* Fenton and Fenton. *Crypzoön proliferum fieldii* is plentiful in the Limeport but scarce in the Allentown Formation. *Archaeozoön* and *Dolatophycus* are probably restricted to the Limeport, although a few doubtful examples of the former were seen in the Allentown.

THICKNESS: The Limeport is about 900 feet thick. No section completely exposes the formation. Along the Bushkill at Easton, the writer calculated the thickness as 950 feet.

CORRELATION AND DATING: The Limeport invertebrates are early late Cambrian or Dresbachian. The formation matches the middle part of the Kittatinny formation in New Jersey, and the lower half of the Conococheague in south-central Pennsylvania. Its tie with the Warrior formation in central Pennsylvania is assumed by the writer on the basis of stratigraphic position and fauna.

Allentown Formation

NAME: The Allentown Formation as defined and named by Wherry (1909) included all strata between his Leithsville below and his Coplay above. As noted, Howell, Roberts, and Willard (1950) separated off the lower part of the formation and named it the Limeport. They retained the name Allentown for the upper part. The topographic expression is usually subdued, but the Allentown has a tendency to form low ridges and cuestas. This property is useful in differentiating areas underlain by the Allentown from those underlain by the Limeport in aerial photographs. Sink holes are more numerous in the Limeport than in the

Allentown. The areal distribution of the Allentown is much the same as that of the Limeport.

LITHOLOGY: The Allentown is lithologically similar to the Limeport, but light- and dark-gray alternations are less pronounced or absent (Pl. 3, fig. 3) The rock is characteristically dark blue gray to light gray but weathers buff or is little changed. The blue stain mentioned in the description of the Limeport is rare in the Allentown, but whitish efflorescence is more common. Texturally, the Allentown is usually dense. Calcarenites, sedimentary breccias, and calcirudites were recognized. Oölite is less abundant than in the Limeport. Shale partings

TABLE 1. INSOLUBLE RESIDUES IN CAMBRIAN AND ORDOVICIAN CARBONATE FORMATIONS OF EASTERN PENNSYLVANIA

Formation names		Hills (1935) (paraphrased)	Knowlton* (Consolidated from sections)	R. H. Lessentine (1952, M.S. thesis, Lehigh Univ.) (Bucks County)
Coplay	Upper	Largely shale, some sand. Large dolocasts common. Chert abundant above middle	..	..
	Middle	..	..	..
	Lower	Many shale particles, little sand or chert	Silt and clay in varying amounts; a few particles of chert. Many rounded, frosted quartz grains	..
Allentown	Upper	Sand with equal amounts of Shale. Some chert and a few dolocasts	Abundant dark-gray to black chert particles and many glassy to frosted, rounded quartz grains at top. Downward, less chert but much dark-gray shale	..
	Middle	..	..	..
	Lower	Shale particles with slightly smaller quantities of sand. Some chert pieces	Many frosted, fine quartz grains; little chert and some dark-gray clay. Black-shale particles may be abundant.	..
Limeport	Upper	Largely sand	Scattered, rounded quartz grains; gray to black clay and a little shale plus gray silt, clay and chert particles	
	Middle	Shale in bottom but sandy toward top Much sand and some shale Equal quantities of sand and shale	..	Limeport as a whole: Rounded to subrounded quartz grains, mostly frosted, and angular quartz grains. Tabular shale fragments and subangular chert. Occasional feldspar grains; mica may be abundant. Among edgewise conglomerates are rounded and subrounded quartz grains, most of them frosted.
	Lower	Sandy top and bottom: shale chips in middle	..	
Leithsville	Upper	Mostly shale with several sand zones. Some "dark" chert, but residues more often "light-colored". More sand in middle	..	..
	Middle	Largely "dark" shale. A few sandy samples. Some chert. Large quantities of rounded quartz grains, some of them secondarily enlarged	..	..
	Lower	"Dark" shale with small quantities of sand and chert	..	..

*Analyses made under the direction of the author. Mr. Knowlton was formerly a graduate student at Lehigh University.

are rare. Some beds contain scattered, clear, well-rounded quartz grains. The bedding ranges from platy to submassive but is mostly uniform, and laminations show on some weathered surfaces. Some bedding surfaces are ripple-marked.

CHEMICAL CONSTITUENTS: Miller (*in* Miller *et al.*, 1939, p. 370) published three analyses of the upper part of his Allentown at Portland, Northampton County, Pennsylvania. Recalculation for the calcite-dolomite ratios gives:

$CaCO_3$	1.1	..	21.2
$CaMg(CO_3)_2$	88.5	94.57	69.7

INSOLUBLE RESIDUES: Hills (1935) published insoluble-residue analyses from the equivalent of the restricted Allentown. They show much sand, shale fragments, and a few cherts and "dolocasts" in the upper part and slightly commoner shale bits, plus sand and some chert, lower. At the author's request, Kenneth B. Knowlton analyzed the uppermost beds along the Bushkill and found rounded, frosted quartz grains, chert, silt, and many dark-gray or black shale fragments. At the Nancy Run section, the lowest beds contained fine-grained, rounded quartz, silt, and pieces of chert and shale. The bottom of the formation on the Bushkill at Easton carries chert spherules and many fragments of black shale. South of Fullerton on the Lehigh River, the uppermost Allentown beds have an abundance of rounded quartz grains and fragments of gray chert, but little shale. Lower, the proportions are reversed with no gray chert but dark-gray shale chips among the rounded quartz grains. G. D. O'Brien also ran tests for insolubles in this section (*in* Willard, 1955a, p. 831). Although not enough insoluble residue analyses have been made to use for correlation, the results are listed for future reference.

FOSSILS: Invertebrate fossils are scarce. Howell (1945, p. 8) reported a Trempealeauan faunule at Portland, Pennsylvania. Hills (1935) cited Foraminifera in pockets. These seem never to have been identified. As with the Limeport, stromatolites are common. Some of these are distinctive of the Allentown. Roberts (*cf.* Fig. 1) illustrated two which he called "Form species A" and "Form species B". Besides these, *Anomalophycus compactus* Fenton and Fenton and *Cryptozoön proliferum fieldii* Fenton and Fenton are known. *Cryptozoön proliferum fieldii* is scarcer than in the Limeport, and *Anomalophycus* and Roberts's two form species are probably restricted to the Allentown.

THICKNESS AND CORRELATION: The Allentown is close to 900 feet thick. Along the Bushkill at Easton the author calculated it at 910 feet. The Allentown correlates with the upper part of the Conococheague of south-central Pennsylvania and the Mines Dolomite of the Bellefonte region of the central part of the State. Because the Kittatinny in New Jersey has not been differentiated, exact matching with it is futile. The Allentown is separated from the Limeport by a disconformity (Pl. 3, fig. 4). The lost interval is approximately Franconian. Above, the Allentown may interfinger with the Coplay ("Beekmantown") of early Ordovician age, as exemplified along the Lehigh River south of Fullerton. An hiatus separates the Allentown from the Coplay at Carpentersville, New Jersey, and in east-central Bucks County the Allentown is absent, so that the Coplay (?) rests on the Limeport Formation.

Lower Ordovician Sedimentary Rocks

When this study was initiated the Lower Ordovician strata were included. However, as the field work progressed it was found advisable to treat them separately. Authorization was given by Dr. H. R. Aldrich (letter of March 26, 1958) to publish a separate account of the Lower Ordovician (Willard, 1958). Because of the relationships of the Lower Ordovician to the Cambrian, the latter is mentioned at several places in the present text. Therefore, it is appropriate to introduce a brief summary of the Lower Ordovician beds in the Lehigh Valley.

"Beekmantown" was for years used to designate the Lower Ordovician of the Lehigh Valley (Miller, 1934; Miller *et al.*, 1939; 1941). The name is inappropriate because of the impossibility of establishing close correlation with the type Beekmantown of northeastern New York, and because the Lower Ordovician dolomitic limestone of the Lehigh Valley transcends the Beekmantown in that it carries Chazyan fossils. The author revived Wherry's (1909) designation, Coplay, and redefined the type locality along the west bank of the Lehigh River near the town of Coplay, Lehigh County. The Coplay includes 1500 feet of dolomitic limestone. It may interfinger with the underlying, uppermost Cambrian, the Allentown Formation, or, where that is absent,

rest disconformably on the Limeport. The Coplay is separated disconformably from the succeeding Jacksonburg of Trenton age. Throughout, the Coplay is characterized by heavy-bedded, light-gray, dolomitic carbonate rocks. The few fossils are those of invertebrate marine animals; no recognizable stromatolites are recorded. Table 1 lists insoluble residues reported from the Coplay.

STRUCTURES

The writer's interpretation of the stratigraphy requires modification of some of the ideas concerning the described structures. Most structures are attributed to the Appalachian revolution and the Taconic disturbance (Willard, 1939; 1943). The stuctures of the Appalachian revolution, superimposed upon the older ones, are not readily distinguished.

The dominant pattern is east–northeast-trending folds and faults. Many of the ridges are asymmetric anticlines. The south flanks may be punctuated by thrust faults. In the limestone valleys, the dolomitic formations have been folded along regional trends. Some writers have emphasized certain spectacular local folds, but the author feels that the complexity of these structures has been overrated. Sections on the banks of the Lehigh River between Allentown and Easton, on U.S. Highway 611 north from Easton and along the Bushkill at Easton, and along the Lehigh River south from Fullerton expose relatively few plications. Instances are available of overturning. The Nancy Run section is in part overturned, as too is that at Fullerton along the Lehigh Valley Railroad.

The writer disagrees with the interpretation on the Northampton County geological map (Miller *et al.*, 1939) that shows "Tomstown" (Leithsville) in the valley north of Bougher Hill. The writer reinterprets the area as a north-tilted fault block in which the formations range northward from Hardyston into Coplay. A prominent fault separates the Coplay from the Precambrian of Elephant Rock. The writer has mapped the trace of this fault into New Jersey, north of Carpentersville, where the Coplay is faulted down against the Cambrian. In the carbonate-floored valley, no trace of Lower Cambrian rocks appeared southeast of Elephant Rock. The valley can be interpreted as a faulted, east-plunging syncline that opens widely into New Jersey.

A well-authenticated thrust strikes westward from the south edge of Phillipsburg into Pennsylvania. On the Delaware, Lackawanna and Western Railroad, opposite Walter Park in the southeastern edge of Phillipsburg, the Upper Cambrian strata are faulted north against the Middle Ordovician fossiliferous Jacksonburg limestone. If the fault trace is projected westward, it aligns with Mt. Parnassus, a small boss of Precambrian gneiss pushed up through younger rocks on the east bank of the Delaware in the south edge of Phillipsburg (*Cf.* Weiler and McCallum, 1955). Farther west in Pennsylvania, a third point on the fault is exposed in a large, abandoned quarry on the Delaware River half a mile south of Easton. Here, the Leithsville is thrust over the Upper Cambrian rocks along a steep south-dipping surface.

The Northampton County geologic map (Miller *et al.*, 1939) indicates an inferred fault connecting Chestnut Hill at Easton with Pine Top and Camels Hump north of Bethlehem. The simplest interpretation is that the Cambrian strata on the south overlie unconformably the Precambrian metamorphic rocks. On the north side there was local faulting so that the Precambrian rocks have been thrust over the Coplay ("Beekmantown" of the county map). Since the county map was published, considerable evidence has appeared to support the interpretation of a fault connecting the Precambrian outcrops on Chestnut Hill with those at Camels Hump. Half way between the outcrops is Green Pond, a large sink hole. Examination of the ground west thereof toward Camels Hump revealed other sink holes in line with Green Pond and Camels Hump. In a patch of woods a mile west-southwest of Green Pond some loose pieces of Precambrian gneiss were found. James Kell, a former graduate student at Lehigh University, ran a magnetometer survey along the supposed fault trace and discovered anomalies. These agree with observations published by Hersey (1944).

CAMBRIAN ENVIRONMENT

Where ecology is empirical and experimental, paleoecology must be inferential and uniformitarian. The paleoecology of Cambrian stromatolites, bioherms, or biostromes in eastern Pennsylvania reveals something of their habits and habitats. From lithology one may attempt to reconstruct the probable environment. That that environment was marine is attested to by the presence of Cambrian marine invertebrate fossils.

Obviously, the area studied holds little promise of contributing to the knowledge of Paleozoic geography of eastern North America. In a description of the paleogeography of eastern North America during Cambrian and early

Ordovician times, Dunbar (1949, p. 129)[3] states:

> At the very outset of the Paleozoic era the Cambrian seas began to encroach upon the lowest lands, and before the close of the period covered more than 30 per cent of the present continent. . . . In these places Cambrian formations attained a great thickness and abound in marine fossils, showing that seas were present here for a very long time. Since the sediments are of types that form in shallow water, it is clear that these places were vast troughs, subsiding slowly as they filled; that is, they were typical *geosynclines*.

Dunbar comments further that the Lower Cambrian beds of the Appalachian geosyncline are mostly detrital, but the rest of the system is limestone and dolomite. The lithologies of the Cambrian and Lower Ordovician rocks of the Lehigh Valley support Dunbar's conclusions. The Lower Cambrian Hardyston Formation is clastic, and the remainder of the Cambrian is dominantly carbonate and shows a progressive decrease in clastic contents.

	Sandstone and shale	*Carbonates*
Allentown Formation	8.2 per cent	91.8 per cent
Limeport Formation	11.0	89.0
Leithsville Formation	25.0	75.0[4]

One may attempt to reconstruct the conditions of origin of the formations studied. The Hardyston was deposited in a shallow sea. Its abrupt vertical and horizontal lithologic and thickness changes, cross-bedding, and ripple marks indicate shifting currents, tidal scour, bars, and barriers. From the relationships of the Hardyston to the Precambrian, one may postulate rapid incursion of the sea. The old land was sometimes buried so quickly that ancient soils were preserved. Thickness variations of the Hardyston from zero to at least 200 feet indicate that the Precambrian surface was irregular. No continuous, basal conglomerate developed such as might be expected if thorough sorting during a more leisurely onlap had occurred.

The passage from the Hardyston sedimentation into Leithsville deposition appears to have been penecontemporaneous throughout the region. The Hardyston (except at one place) grades upward from arenite to shale to limy and sericitic shale into the Leithsville dolomite. Among the dolomite beds of the Leithsville are recurrent mud-cracked shales. Ripple marks indicate agitated water. After the Leithsville Formation was deposited, incursion of detritals further decreased, and stromatolites appeared. Among the clastic materials found in the insoluble residues of the Limeport and Allentown are rounded, frosted quartz grains that could be of dune origin. A hiatus separates the Limeport Formation from the Allentown.

Where, in relationship to the land, were all the Cambrian carbonate formations deposited? Several possibilities come to mind. It has been shown that, except for an occasional shale bed, the commonest detrital material in the Cambrian carbonate rocks is siliceous sand. Many of the grains are rounded, spherical, and frosted. Commonly, these grains are scattered rather than in beds, although both have been observed. It is reasonable to believe that little or no water-borne sediment was being introduced from the land, but that wind-formed sand grains were. These may have been derived from seashore dunes, but may have been blown seaward from farther inland. The coast might have been so remote that little clastic material reached the locale. An arid climate could explain the absence of fluviatile material, except in limited amounts, and account for the eolian sand grains. A terrane of low relief with sluggish streams might account for the absence of detritus. Finally, but only remotely possible, the land might have been largely composed of carbonate rocks.

Considering the invertebrate fossils and the lithology, one may conclude that the organisms that formed the stromatolites lived in shallow, agitated, marine water in which little mud or sand accumulated. Detailed sections[5] (sections A-F) support the validity of Link's (1950) postulated conditions for a transgressive sea: a basement beneath the sedimentary rocks, no evaporates, no porous reef rock, no basinward pontic black or gray shales such as accompany offlap.

A detailed section (section G) of a portion of the Limeport illustrates the relationships of

[3] Quoted with permission of the publishers, John Wiley and Sons, and of the author.

[4] Estimates based upon scattered information; no complete sections available.

[5] Deposited as Document number 6895 with ADI Auxiliary Publications Project, Photoduplication Service, Library of Congress, Washington 25, D. C. A copy may be secured by citing the Document number and by remitting $3.75 for photoprints, or $2.00 for 35-mm microfilm. Advance payment is required. Make checks or money orders payable to: Chief, Photoduplication Service, Library of Congress.

the bioherms to their environment. It was recorded at cuts 4 miles east of Bethlehem, on the north side of the Lehigh River, along a spur track of the Central Railroad of New Jersey. Only significant thicknesses are given.

From the writer's sections and other, scattered observations, certain generalities of habit and habitat may be deduced for the stromatolites:

Heads grew disconformably on underlying beds of whatever lithology	92 per cent
Heads grew conformably to underlying beds of whatever lithology	8 per cent
Tops of heads disrupted and more or less eroded (Pl. 5, fig. 2)	31 per cent
Tops of heads preserved entire, of which about 50 per cent are shale-smothered (Pl. 5, figs. 3, 4)	69 per cent

Authors seem to agree that algal bioherms may form on either firm or unconsolidated bottoms. Black (1933) noted that deposits produced by living algae pass laterally into deposits that were not produced by algae. These relationships are illustrated by the Cambrian forms noted herein. Bradley's (1929) observations on the association of algae with oölites are significant, even though applied to a freshwater environment. Certainly, the Upper Cambrian rocks studied here, particularly the Limeport, demonstrate the association of algae with oölites. Ginsburg (1955) has observed in Florida living carbonate-depositing marine algae that strikingly resemble, superficially or otherwise, the Cambrian forms. He reports the algae as living in shallow, protected marine waters in open, intertidal zones where the flats may be flooded at high water and stratigraphic laminations may develop.

Evidently, the Cambrian organisms preferred to grow on clean bottoms from which they were separated by a distinct surface of parting. Fewer grew on a surface from which they are not clearly separated. In the first instance, some consolidation of the sediments may have taken place before the stromatolites "took root", in the second, nonconsolidation of bottom sediments is assumed. The proportions of those heads that were buried entire to those that have the tops disrupted is about 2:1. Because those stromatolites with entire heads are as often as not shale-covered, the organisms were presumably smothered by a fatal influx of mud. Those in which the heads were broken were eroded by wave or current movements. The fragments of many of the broken heads were incorporated in the succeeding beds, indicating that the transportation of those fragments was limited to the immediate neighborhood.

Carozzi (1953) published significant facts on the relationships between carbonate-secreting algae and the precipitation of carbonates. During photosynthesis finely divided precipitate of carbonate may form over the thallus. The pH and alkalinity of the water may vary between daylight and night hours, as CO_2 may be removed or added. Also, $Mg(OH)_2$ can be deposited with change of environmental composition. With the eventual change of this $Mg(OH)_2$ to carbonate, a possible clue to the formation of dolomite or dolomitic deposits is suggested.

SUMMARY

The Cambrian sedimentary rocks in eastern Pennsylvania are about 3000 feet thick. They consist mostly of dolomite, but the lowest 200 feet or so are detrital-derived. This basal formation, the Hardyston, varies lithologically and in thickness and might have been laid down in a sea advancing rapidly over an old surface that was long subjected to subaerial weathering and erosion. The carbonate rocks that follow, the Leithsville, the Limeport, the Allentown, and probably the Lower Ordovician Coplay, show a progressive decrease in clastic materials. All the Cambrian formations except the Leithsville contain marine invertebrate fossils; the Limeport and Allentown have abundant stromatolite remains, absent in the Coplay.

Possibly changes in salinity, temperature, depth, and/or detritus contents wrought the changes in fauna and flora. No immediate explanation appears for the barrenness of the Leithsville except perhaps its higher clastic content. Elsewhere in adjacent North America the Lower Ordovician sedimentary rocks have many "cryptozoöns" (Bassler, 1919). None have been identified from the Coplay. That formation does carry a few large, marine gastropods. Today, large snails may be found as benthonic dwellers in relatively deep ocean waters. Is there a corollary? If the waters were deep, precluding light penetration, the absence of algae is explainable.

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UNIVERSITY OF MICHIGAN

Source of Soundings :-
Royal Australian Navy Hydrographic Office
Royal New Zealand Navy Hydrographic Office
British Admiralty Charts

PS43 Limits of R.A.N. coastal charts

2770 2780 2790 Echo soundings

•2730 Single soundings

NT7 North Tasman seamount No.7

137 Shoalest depth to guyot in fathoms

Depth in fathoms ; Contour interval : 500 fathoms
Scale: 1:1,000,000 at Lat. 46°

150° 30' 151° 30' 152° 30' 153° 30' 154° 30' 155°

30° 30' 31° 30' 32° 30' 33° 30'

Coffs Harbour

Smoky Cape

Port Macquarie

Sugarloaf Point

Newcastle

Profile 5

PS 51 PS 50 PS 55 PS 56

•1850 •2572 •2770 •2565 •2246 •3012 •2740 •2725 •2998 •2795 •1613 •3250 •2608 •1867 •2647

NEW SOUTH WALES

152° 30' 153° 30' 154° 30' 155° 30' 156° 30' 157° 30' 158° 30' 159°

Coffs Harbour

Smoky Cape

Port Macquarie

Sugarloaf Point

fathoms

Office

Derwent Hunter Guyot

Barcoo Guyot

Taupo Bank

Profile 1

Profile 2

Profile 4

Profile 5

NT 7

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P.S. 55

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Coffs Harbour

Smoky Cape

Port Macquarie

Sugarloaf Point

fathoms

Office

Derwent Hunter Guyot

Barcoo Guyot

Taupo Bank

Profile 1

Profile 2

Profile 3

Profile 4

NT 7

PS 50

PS 51

PS 55

PS 56

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156° 30' 157° 30' 158° 30' 159° 30' 160° 30' 161° 30' 162°

Middleton Reef

Elizabeth Reef

Derwent Hunter Guyot

Barcoo Guyot

Taupo Bank

Profile 1

Profile 2

Profile 3

NT 7

Lord Howe Is.

Ball's Pyramid

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Ball's Pyramid
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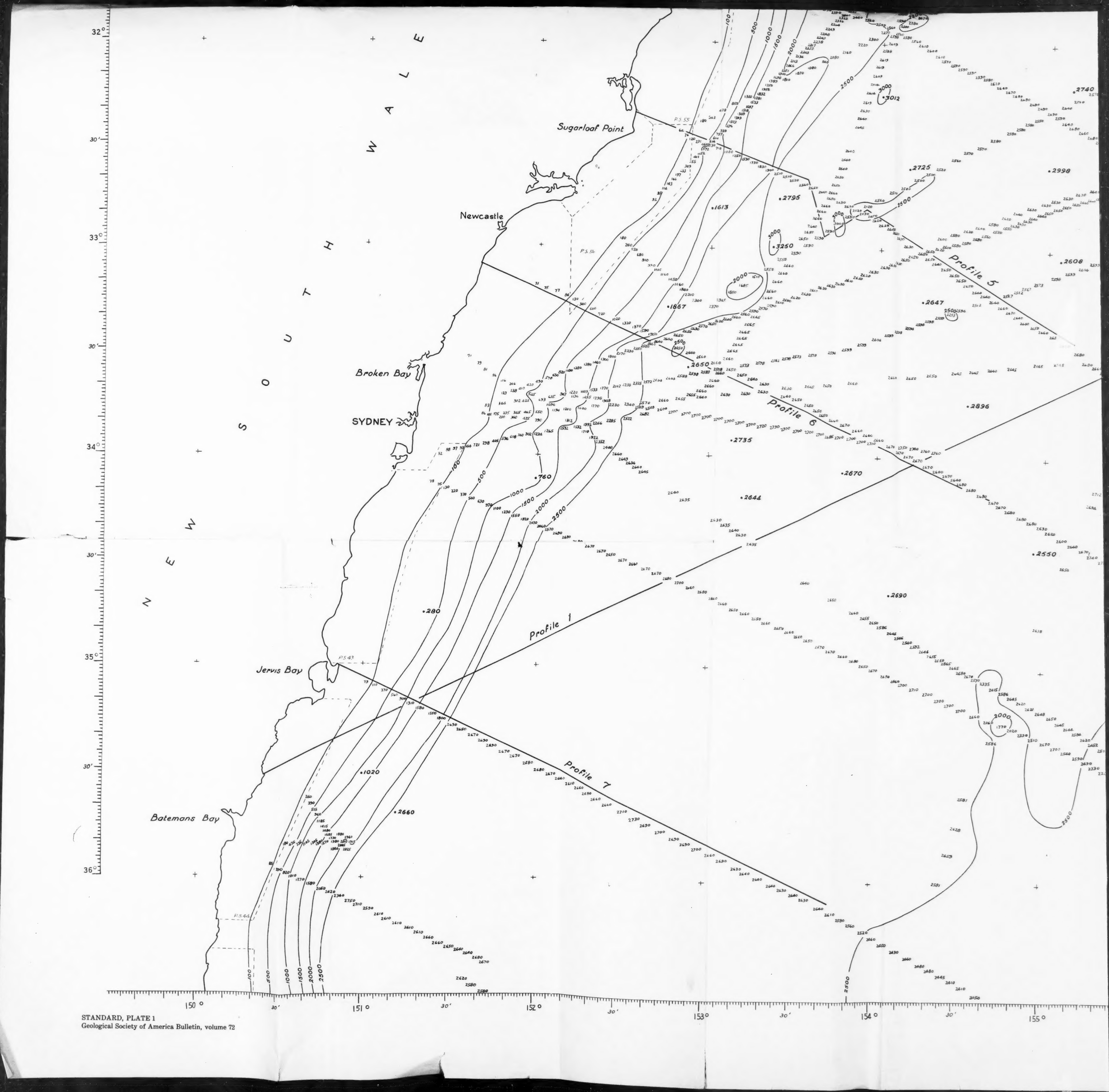

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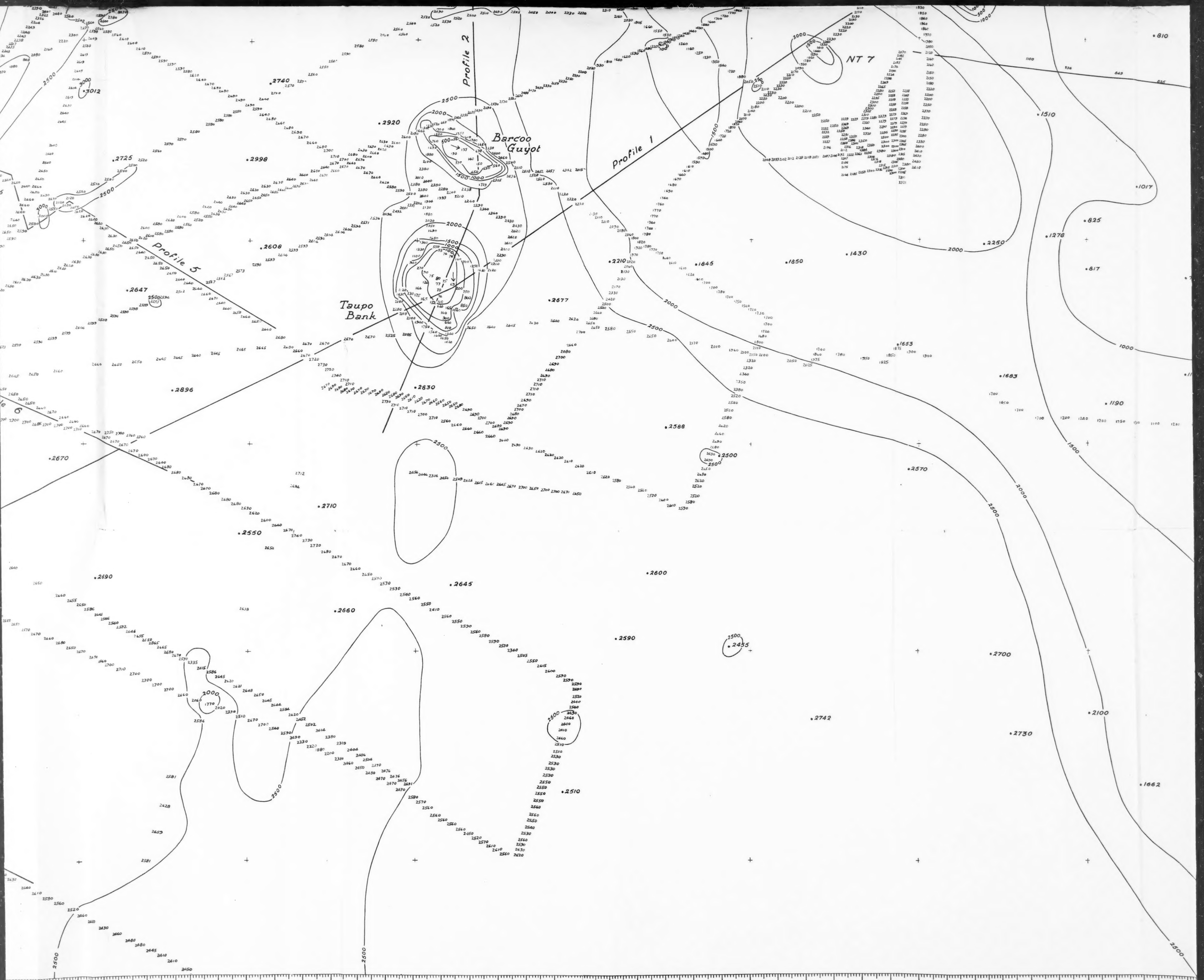

BATHYMETRIC MAP OF THE MIDDLE PART OF THE TASMAN SEA

NT 7

Profile 3

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1653 1683 1190 1160 946 1318 933 1666 508 493 518 678 646 858 1000 870 667 496 1110 500

2570 2500 2000 2700 2742 2730 2100 1442 1760 1662 1642 1792 1690 1643 1050 650 1000 1500 2000 2500 850 674 702 646 625 630

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159° 30′ 160° 30′ 161° 30′ 162° 30′ 163° 30′ 164° 30′

N SEA

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Submarine Geology of the Tasman Sea

Abstract: The physiographic features of the continental margin of eastern Australia, the Tasman Basin, Lord Howe Rise, and the Coral Sea Platform are described and discussed geologically. Three guyots, each having more than 14,000 feet of relief and a platform depth of less than 150 fathoms, are mapped and described.

The present continental slope of southeastern Australia west of the Tasman Basin marks the maximum eastward development of the Australian continent. Lord Howe Rise is considered orogenic in origin and probably of Early Paleozoic age. The Tasman Basin is a stable area underlain by permanent ocean-type crust which may have acted as a nucleus for the eastward growth of the island arcs which lie between the Tasman Basin and the South Pacific Basin.

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INTRODUCTION

The geology of the southwest Pacific, including the Tasman Sea, received much attention around the turn of the century (Suess, 1906; Gregory, 1910; Marshall, 1911). Benson (1923; 1924) gives an excellent review of many of these early papers and lists more than 400 references.

Most early writers believed in the existence of an ancient Australian continent extending eastward from the present Australian coast to the South Pacific Basin including both Fiji Islands and New Zealand. Among the more recent authors who have expressed a similar view are Bryan (1944), Macpherson (1946), de Jersey (1946), Gill (1952), Glaessner (1952), Gutenberg and Richter (1954), Marshall and Narain (1954), Fleming (1957), and Gill (1958). Hess and Maxwell (1949) indicated no crustal thickness but suggested that the island arcs of the southwest Pacific are the result of progressive outward growth of the Australian continent from a nucleus in Western Australia.

Officer (1955) found that there was no continental-type crust in the southwest Pacific and the crustal thickness of the Tasman Basin was only 5 km, the same as that of the South Pacific Basin. Geological interpretation of recent oceanographic surveys conducted by the Royal Australian Navy supports Officer's conclusion.

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BATHYMETRY

Until recently there has been little oceanographic knowledge of the Tasman Sea, most of which was based on pre-1900 single, lead-line soundings many of which were several hundred miles apart. Recent work by the Royal Australian Navy has made it possible to construct a bathymetric map of part of the Tasman Sea based on continuous echo-sounding profiles (Pl. 1). The portion of the Tasman Sea which has been mapped in most detail is included in the world-wide network of "Plotting Areas for Ocean Soundings" 413 and 414. However, detailed information is not available for most of the Tasman Sea, and the maps presented here are based upon information available before June 1961.

The names of physiographic features used in this paper are those used by Brodie (1952;

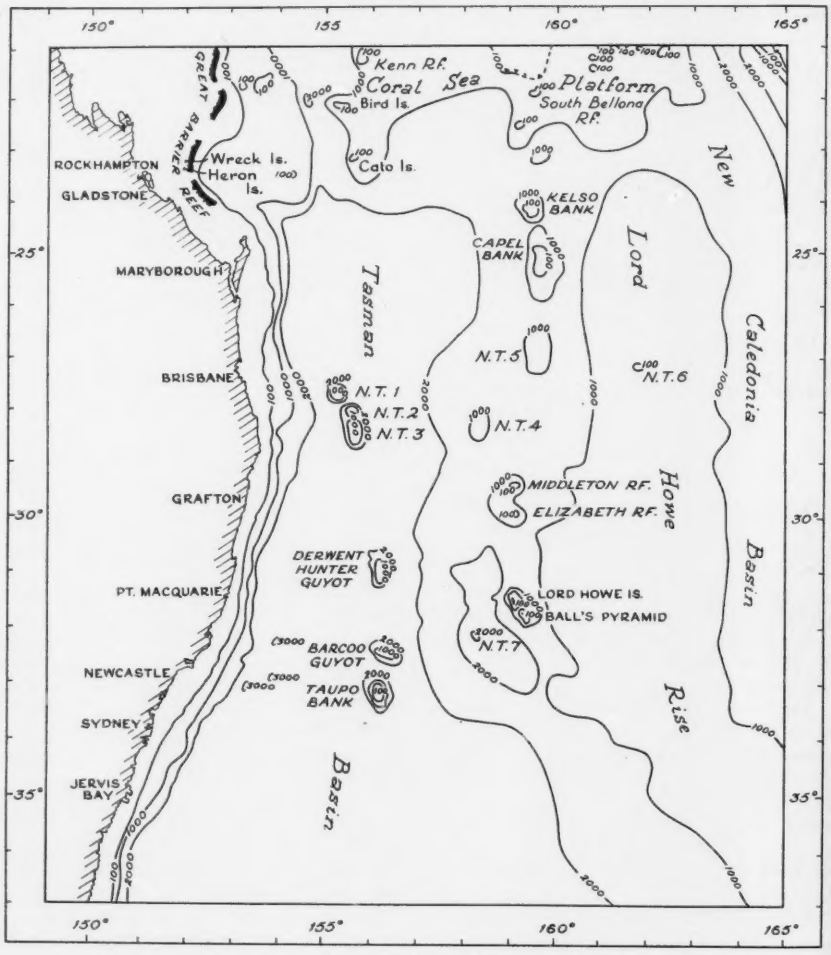

Figure 1. Location map of the physiographic features of the Tasman Sea

1958) and Wiseman and Ovey (1955). The proposed new names follow the rules suggested by Wiseman and Ovey (1953). The name "Coral Sea Platform" is proposed for the area north of the Tasman Basin.

Three guyots have been mapped. The southern one, the Taupo Bank, was named by Fleming and Brodie, (1951, unpub. ms., New Zealand Oceanographical Committee, Wellington, N. Z.). The names Derwent Hunter Guyot and Barcoo Guyot are proposed for the other two. The DERWENT HUNTER is a research schooner belonging to the Australian Commonwealth Scientific Industrial Research Organization which discovered the northern guyot in 1958. The BARCOO is a Royal Australian Navy survey ship which discovered the central guyot in 1956.

Seven seamounts have been given the names NT 1 (North Tasman seamount 1) through NT 7. The seamount NT 1, NT 2, and NT 3 were discovered in 1902 by the C. S. BRITTANIA while laying cable between Brisbane and Norfolk Island. David (1932) referred to these as the Tasmantides Volcanoes.

PHYSIOGRAPHIC FEATURES

Continental Margin

The continental shelf between the Tasman Sea and the Australian continent ranges in width from 40 miles in the north near Maryborough to 8 miles in the south near Jervis Bay (Fig. 1). This is noticeably less than the 200 miles of continental shelf near the southern end of the Great Barrier Reef. The average inclination of the continental shelf along the Tasman Sea is about 0°08′, and the abrupt change in slope, the shelf break, occurs at about 55 fathoms. These figures are similar to those recorded by Shepard (1948a) and Dietz and Menard (1951) for other areas. The average continental slope of the Australian continent along the Tasman Sea is about 6°, but further soundings will probably reveal steeper slopes and scarp lines. As noted by Shepard (1959) the continental slope off southeastern Australia may be classed as complex; in many places the lower portion of the slope is much steeper than the upper. The steepness of the lower slope is 15° in many places. This type of slope, typical of those leading into deep trenches, is also found off the coast of Japan and south of the Aleutian Islands.

A wider continental shelf is found off the mouth of the Hawkesbury and Hunter rivers, probably due to an increase of sediments deposited by the rivers. A definite delta-type deposit is shown on the coastal charts of this area at a depth of 50 fathoms.

Wyrtki (1960) noted the eastward swing of the southeast Australian current away from the continent at about 32° S. latitude. The narrowness of the continental slope and shelf north of 32° S. latitude is possibly due to this current which has prevented the accumulation of sediments, in much the same way as the Gulf Stream has off the east coast of Florida (Shepard, 1959).

An 1890-foot well drilled in 1959 on Wreck Island, in the southern part of the Great Barrier Reef, passed through 530 feet of Recent coral-reef material. Tertiary sediments were penetrated from 530 feet to 1795 feet (Mott, 1960). The underlying bedrock is probably Paleozoic metamorphic rocks similar to those on the mainland. A 732-foot well drilled on Heron Island, 6 nautical miles south of Wreck Island, revealed a marked increase in terrigenous material at 500 feet. This was interpreted as representing the base of Recent coral material and was thought to have been associated with the eustatic change of sea level of 300 feet during the last Pleistocene glacial period (Richards and Hill, 1942). This would indicate a post-Pleistocene subsidence for the continental shelf of about 200 feet.

No well-defined submarine canyon has been found off the coast of Australia, probably because of a lack of soundings rather than a lack of canyons. Most of the coast of southeastern Australia has been surveyed by echo-sounding methods to a depth of 100 fathoms, but these soundings were made along widely spaced traverses at right angles to the coast, and small canyons may have been missed. If such canyons are found they will probably be the offshore type (Kuenen, 1950) located along the continental slope at a depth of more than 100 fathoms. Indications of possible canyons were found off the mouth of the Hawkesbury River, off Smoky Cape, and 12 miles northeast of Coffs Harbour. The Tectonic Map of Australia, published by the Australian Bureau of Mineral Resources, shows two submarine canyons off the coast: one is off the eastern end of Bass Strait, the narrow body of shallow water (Jennings, 1959) separating the mainland from the island of Tasmania, and the other is off the mouth of the Swan River in Western Australia. However, these submarine canyons, like those of Sprigg (1947) off South Australia and

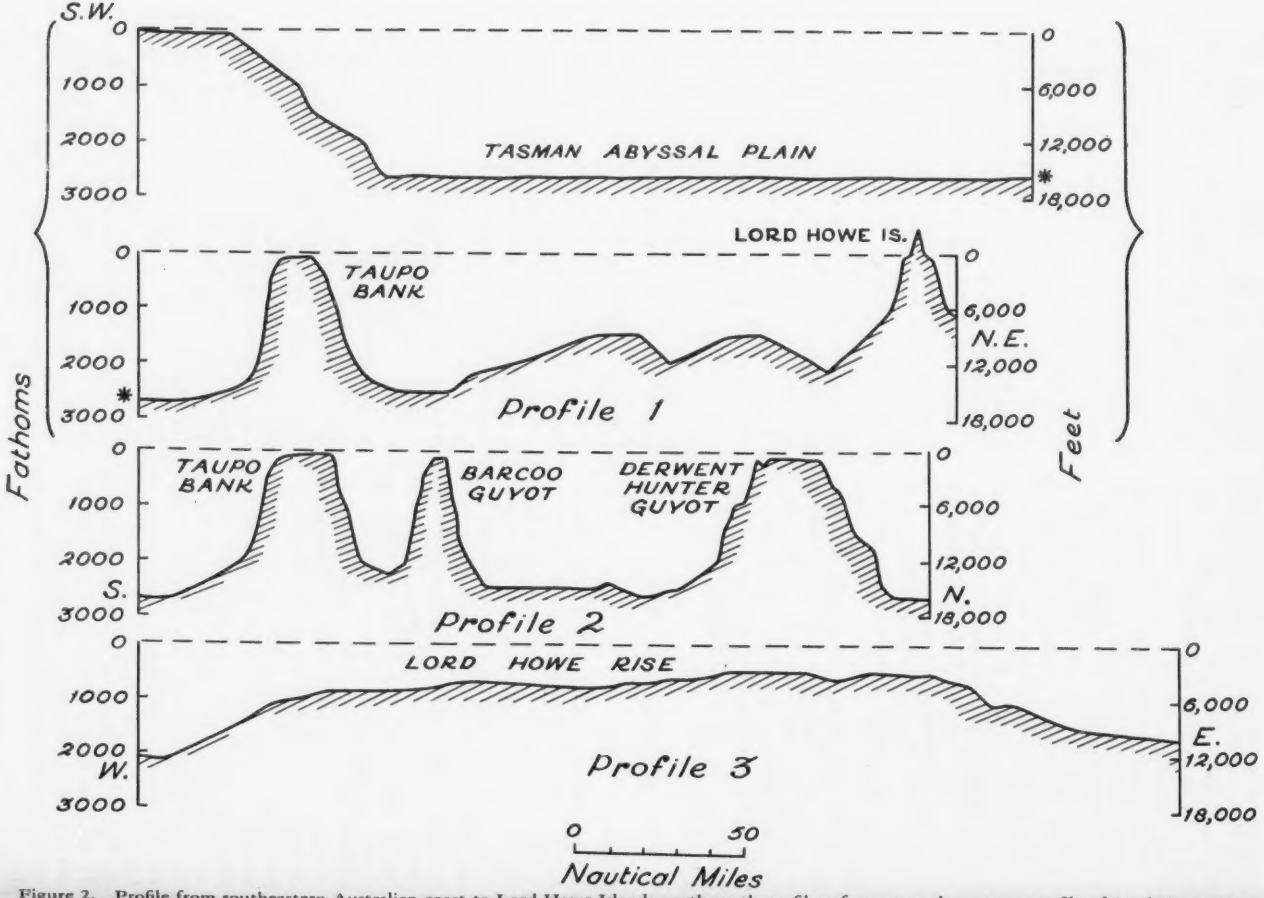

Figure 2. Profile from southeastern Australian coast to Lord Howe Island; north-south profiles of guyots and east-west profile of Lord Howe Rise

Shepard (1948b) off Shark Bay in Western Australia, are based on few soundings, and their existence is not certain.

Tasman Basin

The International Nomenclature Committee on ocean-bottom features (Wiseman and Ovey, 1955) gives the boundaries for the Tasman Basin as the Australian continent on the west, the Lord Howe Rise on the east, and the Macquarie Rise on the south. No boundary was given on the north so that, in the present paper, the Coral Sea Platform (Fairbridge, 1950) is considered as the northern boundary of the Tasman Basin, as well as the dividing line between the Coral and Tasman seas.

The floor of the Tasman Basin is characterized by a broad abyssal plain at a depth of about 2650 fathoms (Fig. 2, profile 1) and becomes more irregular where the basin narrows near the Derwent Hunter Guyot (Fig. 3, profile 4). Three places with depths of more than 3000 fathoms have been found a few miles east of the base of the continental slope. Echo soundings indicate that one deep is located in a narrow trench 2460 feet deeper than the surrounding ocean floor (Fig. 3, profile 5).

Three guyots, the Taupo, the Barcoo, and the Derwent Hunter with heights of more than 14,000 feet have been found 160–200 nautical miles east of Australia. The shoalest depth of the submerged platform is 69 fathoms (414

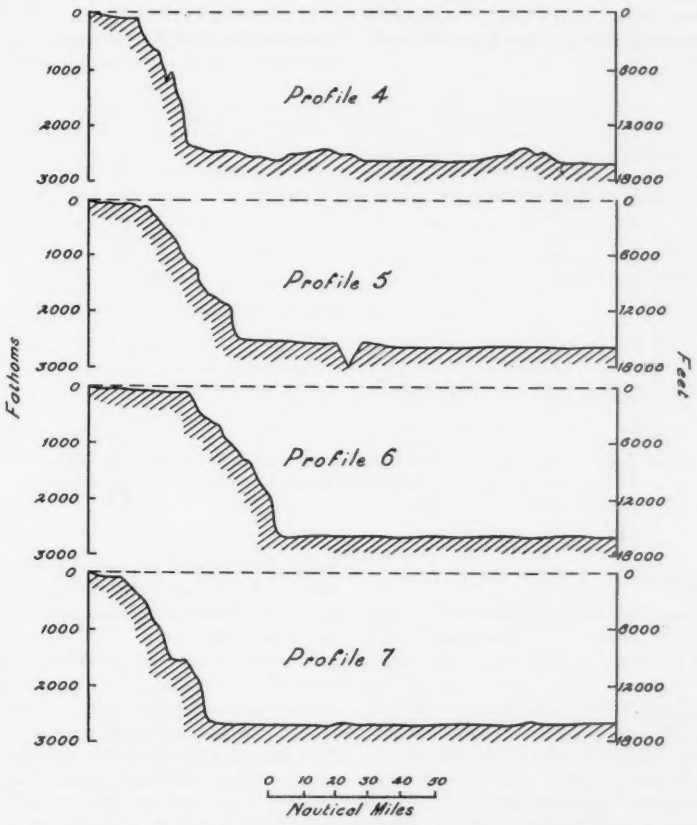

Figure 3. Profiles of the continental shelf and slope of southeastern Australia

Figure 2. Profile from southeastern Australian coast to Lord Howe Island; north-south profiles of guyots and east-west profile of Lord Howe Rise

Nautical Miles

feet) for the Taupo, 137 fathoms (822 feet) for the Barcoo, and 150 fathoms (900 feet) for the Derwent Hunter (Fig. 4; Table 1). The average slope of the guyots is about 18° but ranges from 10°–20°.

About 100 nautical miles southeast of Brisbane three seamounts, NT 1, NT 2, and NT 3, have heights of 11,430 feet, 13,620 feet, and 12,570 feet respectively. These are possibly guyots, but no detailed survey has been made of the area.

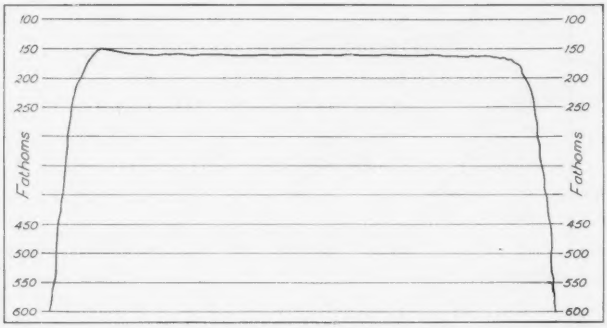

Figure 4. East-west profile of Derwent Hunter Guyot, width of platform 11.2 nautical miles, taken from echogram made by H. M. A. S. WARREGO, May 13, 1960

Lord Howe Rise

Lord Howe Rise, a broad asymmetrical ridge 100–200 nautical miles wide, is bounded by the Coral Sea Platform on the north, the northwest part of South Island (New Zealand) on the south, Tasman Basin on the west, and New Caledonia Basin on the east. The crest is generally about 700 fathoms deep, but 180 nautical miles southeast of Lord Howe Island a large area is less than 500 fathoms deep (Pl. 1). The base starts at a depth of about 2500 fathoms. In places, such as 40 nautical miles southwest of Lord Howe Island, foothills are developed near the base of the ridge (Fig. 2, profile 1).

Lord Howe Island, on the west flank of Lord Howe Rise, reaches a height of 2838 feet above sea level, is crescent-shaped and approximately 1 nautical mile wide and 7 nautical miles long. It is located in the center of the wave-cut platform which is much larger than the island (14 by 20 nautical miles) and has a depth of about 50 fathoms. This platform was presumably formed by wave action during a eustatic change in sea level associated with Pleistocene glaciation.

Balls Pyramid, a pinnacle of basalt 12.5 nautical miles south of Lord Howe Island, rises to a height of 1811 feet. The pinnacle, which is only 0.2 mile wide at its base is in the center of a wave-cut platform several miles wide. The wave-cut platforms surrounding Lord Howe Island and Balls Pyramid are separated by more than 2000 feet of water and probably were never connected. Both Lord Howe Island and Balls Pyramid are located near the west flank of Lord Howe Rise in ocean depths of more than 2000 fathoms on the west and 1100 fathoms on the east.

Elizabeth Reef and Middleton Reef, 130 nautical miles north of Lord Howe Island and about 5 nautical miles wide, rise out of 1000 fathoms on the west and are separated from the crest of Lord Howe Rise by 1500 fathoms on the east. Capel Bank, Kelso Bank, NT 4, and NT 5 also on the western flank of Lord Howe Rise, rise out of 1400 fathoms on the west and have shoalest depths of 10 fathoms (Kelso Bank), 25 fathoms (Capel Bank), 581 fathoms (NT 4), and 158 fathoms (NT 5). NT 6, near the crest of Lord Howe Rise has a shoalest depth of 13 fathoms (Table 1).

Coral Sea Platform

The Coral Sea Platform, an east-west ridge extending from Kenn Reef, Bird Island, and Cato Island toward New Caledonia, is less than 1000 fathoms deep and is covered by scattered coral reefs, sand cays, and coral islands. It is

separated from the continental slope by a narrow trough more than 2000 fathoms deep. Lord Howe Rise joins the Coral Sea Platform 120 nautical miles northeast of Kelso Bank, at a depth of about 1250 fathoms.

GEOLOGICAL INTERPRETATION

Tasman Basin

The idea has persisted for many years that the eastern margin of the Australian continent borders the South Pacific Basin near Marshall's Andesite Line and that the Fiji Islands and New Zealand are part of the Australian continental mass. Schuchert (1916) and Browne (1947) considered the eastern margin of the Tasman geosyncline, parallel to the eastern edge of Australia, to be in a fixed position in the vicinity of Lord Howe Rise, and Glaessner (1952) states that this has been generally accepted. Gill (1952) stated that the Tasman geosyncline and New Zealand geosyncline developed as parageosynclines of the large Austzealandic continent which stretched south to Antarctica, and again in 1958 (p. 107) Gill stated that "geologically, New Zealand has been regarded as the eastern edge of a great continent which includes Australia and New Zealand as well as the area between them." Glaessner (1952, p. 113) stated that "There is certainly no indication of the existence of an arm of the sea along the coast of eastern Australia." Gutenberg and Richter (1954), while recognizing the occurrence of an extremely stable oceanic area east of Australia, considered that the highly active circum-Pacific Belt including the Solomon Islands and New Zealand marks the edge of the Australasian continental area. Thus, the Tasman Basin is viewed as part of the Australian continent which (Glaessner, 1952, p. 114), "has subsided since Tertiary time" and, "as a unified geomorphological feature . . ., is of late or post Tertiary origin."

Officer (1955) and Eiby (1958) showed that no continental-type crust exists east of the present Australian coast. The crust above the M discontinuity along the eastern coast of Australia is estimated by Doyle, Everingham, and Hogan (1959) to be 37 km thick. The continent is separated from the 5-km thick crust of the Tasman Basin by the steep, complex continental slope. The same type of slope, showing steepening of the lower slope, is adjacent to the Aleutian Trench and Japanese Trench and has been compared with the continental slope of southeastern Australia (Shepard,

TABLE 1. GUYOTS, SEAMOUNTS, BANKS, AND REEFS OF THE TASMAN SEA

Name	Lat. S.	Long. E.	Height (Feet)	Minimum depth (Fathoms)	Platform dimensions (Nautical miles)
			Guyots		
Derwent Hunter	30–50	156–15	14,136	150	16.0 x 11.2
Barcoo	32–35	156–15	14,178	137	14.5 x 4.0
Taupo Bank	33–10	156–08	14,586	69	15.0 x 8.1
			Seamounts with undetermined summit shape		
NT 1	27–44	155–15	11,430	495	
NT 2	28–02	155–36	13,620	220	
NT 3	28–25	155–38	12,570	228	
NT 4	28–28	158–22	5,630	581	
NT 5	26–50	159–34	7,870	158	
NT 6	27–10	161–50	5,320	13	
NT 7	32–06	158–25	3,540	1,410	
			Banks		
*Kelso Bank	24–09	159–25	8,060	10	
*Capel Bank	25–06	159–38	8,150	25	
			Reefs of the Tasman Sea		
Middleton	29–28	159–06	6,000	Sea level	5.0 x 3.2
Elizabeth	29–56	159–04	6,000	Sea level	4.7 x 3.0

*Probably guyots

Officer (1955) and Eiby (1958) showed that no continental-type crust exists east of the present Australian coast. The crust above the M discontinuity along the eastern coast of Australia is estimated by Doyle, Everingham, and Hogan (1959) to be 37 km thick. The continent is separated from the 5-km thick crust of the Tasman Basin by the steep, complex continental slope. The same type of slope, showing steepening of the lower slope, is adjacent to the Aleutian Trench and Japanese Trench and has been compared with the continental slope of southeastern Australia (Shepard,

1959). Voisey (1959b) has compared the Permian of the New England area of eastern Australia with the present Japanese Islands. Because the deepest part of the Tasman Basin lies along the western margin near the base of the continental slope, these deeps are possibly remnants of an ancient southeastern Australian trench which is nearly filled with sediments and analogous to the Japanese and Aleutian trenches at the margins of continents. Dooley (1959) and Dooley and Goodspeed (1959) in an underwater gravity survey of the Great Barrier Reef noted a rise in gravity readings eastward toward the edge of the continental shelf. This is typical of marginal areas between continental and ocean-type crust. Two of the gravity traverses, both in the southern part of the Great Barrier Reef near Wreck Island, showed a negative reversal of the gravity gradient. This negative anomaly may be related to a sediment filled trench.

In this view, the deeps of the Tasman Basin are considerably older than previously thought and may represent the remains of the tectogene (foreshore deeps) associated with orogenic continental growth. Hess and Maxwell (1949) and Hess (1955) stated that progressive outward migration of belts of orogeny associated with continent growth can be identified by characteristic belts of serpentinized peridotite which parallel the tectogene axis and are associated with early stages of island-arc orogeny. Such serpentinite belts (of Permian age, Voisey, 1959a) are present along the eastern coast of Australia. According to Hess (1960) it is not essential that island-arc deformation represent early stages of Alpine-type mountain building; both island arcs and Alpine-type mountains represent different kinds of reaction to the same type of tectonic force,—one developing on an ocean-type crust, and the other on a thin marginal-type continental crust.

In Australia, the eastward growth of the continent from the Western Australian Precambrian shield can be traced by parallel orogenic belts (Voisey, 1959a). This type of accretionary growth of continents was suggested by Wilson (1949), Kay (1951), and Poldervaart (1955), and the Tasman geosyncline along the eastern margin represents the last stage of this development in Australia. The deeps of the western Tasman Basin may represent the remains of the tectogene associated with the development of the Tasman geosyncline, and the present continental slope of southeastern Australia represents the maximum eastward growth of this geosyncline. The Tasman Basin, therefore, is not, and never has been, a part of the Australian continent.

The exact nature of the 5-km thick crust of the Tasman Basin is not known. It is, however, of the same type as that of the South Pacific Basin (Officer, 1955). This type of crust underlying permanent ocean basins was called Thalasso-craton by Fairbridge (1955) and is considered the oceanic counterpart of Kay's (1951) hedreocraton, which is the area of semi-rigid crust on continental masses. Hess (1960) has suggested that the so-called basaltic crust of ocean basins may be peridotite two-thirds serpentinized so that the "crust" is, in reality, altered mantle.

Seamounts and Guyots

Seamounts and guyots are common in the Tasman Basin. Many of them are roughly aligned, in a manner common in the North and South Pacific basins. Hess and Maxwell (1949) attribute this to faults or shear zones in the crust.

Taupo Bank, Barcoo Guyot, and Derwent Hunter Guyot are volcanoes, probably similar to the olivine-rich alkali basaltic volcanoes of the Pacific Basin and the volcanoes of the Tertiary alkali olivine basalt province of eastern Australia (David, 1950), which were truncated by wave action. Since the period of beveling, the guyots have presumably sunk to their present position. If the Taupo Bank, with its 414-foot platform, were beveled during the sea-level change associated with Pleistocene glaciation, only about 100 feet of downwarping would be required to bring it to its present position. Drill data from the southern part of the Great Barrier Reef indicate a subsidence of about 200 feet of the continental shelf since Pleistocene. Volcanic material was dredged from NT 2 at a depth of 527 fathoms (David, 1932).

Volcanic Islands and Reefs

Lord Howe Island and Balls Pyramid are composed of olivine-rich alkali basalt, as are most of the islands of the North and South Pacific basins. Both islands have broad, flat wave-cut platforms much larger than the present islands (Standard, manuscript in preparation). Flint (1957) estimated the maximum drop of sea level during Pleistocene glaciation was about 300 feet (50 fathoms). These platforms, located at a depth of about 50 fathoms, are therefore considered to have been formed

by wave action during this period of lowered sea level.

The coral reef along the west side of Lord Howe Island is the farthest position south (32° S. latitude) at which coral reefs have been recorded. The reef has grown on top of the wave-cut platform and is therefore late Pleistocene to Recent. However, true reef-building coral of the *Pavona* type have recently been found in Sydney harbor (34° S. latitude).

It appears reasonable to assume that Elizabeth and Middleton reefs were formed by growth of coral reefs on top of guyots, beveled during the Pleistocene. Kelso Bank and Capel Bank are table reefs with minimum depths of 10 and 25 fathoms, probably formed in the same way except that coral growth was too slow to reach the surface because of either the rise in sea level, or the subsiding of the guyots.

A sample of branching-type coral, probably *Acropora*, was dredged from the Derwent Hunter Guyot at a depth of 196 fathoms (1164 feet) (R. M. Davies, personal communication). This coral could not have grown at such a depth; the guyot has, therefore, subsided several hundred feet.

Lord Howe Rise and Coral Sea Platform

Lord Howe Rise has a crustal thickness of 20 km and is part of the orogenic belt which, with New Zealand, has been built by successive orogenies upon an ocean-type crust (Officer, 1955). The topographic change associated with the change from the thin stable ocean crust to the thicker crust associated with orogenic activity is shown by the abrupt transition from the abyssal plain to the "foothills" of the Lord Howe Rise (Fig. 2, profile 1).

Glaessner (1950) considered Lord Howe Rise to be part of the inter-Melanesian arc system and suggested that structurally it is related to southeastern New Guinea via Rennell Island, in the southern part of the Solomon Islands. The writer believes it is probably not related to southeastern New Guinea via Rennell Island because it joins the Coral Sea Platform in the north, and there is no suggestion that it continues to Rennell Island.

In the south Lord Howe Rise joins the northwest part of South Island, New Zealand. Brodie (1952) believes that Lord Howe Rise is an anticlinal structure and its continuation on the eastern side of New Zealand, the Campbell Platform, has been offset 300–400 miles at the Kermadec Trough. Brodie (1958) considers Lord Howe Rise to be at least early Tertiary. Kingma (1959) considers it a continuation of the structure at the northwestern part of South Island. Grindley, Harrington, and Wood (1959) show that most of the formations in the northwestern part of South Island are Cambrian and Ordovician; these formations may extend to the northwest along Lord Howe Rise. If the structures and sediments of Lord Howe Rise and northwest South Island are the same age, Lord Howe Rise may be as old as Early Paleozoic. It may then be the oldest of a series of parallel island arcs which become progressively younger eastward to the present active circum-Pacific Belt. Hess and Maxwell (1949) are essentially correct in their interpretation that the parallel orogenic belts become younger eastward, except that the area between Lord Howe Rise and the circum-Pacific Belt becomes a unit separated from the Australian continent which originated with the Tasman Basin as a nucleus (Thalasso-craton) rather than the Western Australian shield.

The Coral Sea Platform certainly appears to be more closely related to the orogenic activity of the Lord Howe Rise than to the stable Tasman Basin. Fairbridge (1950) suggested that is is a part of the subsided continental land mass which forms a northward-tilted block, but this is probably not true because the crustal thickness in the area is only 15–20 km.

CONCLUSIONS

The continental slope of southeastern Australia west of the Tasman Basin marks the maximum eastward development of the continent.

An ancient southeastern Australian Trench, now nearly filled with sediments, may exist along the western part of the Tasman Basin near the base of the continental slope.

The floor of the Tasman Basin which has a crustal thickness of 5 km, is a very stable area, topographically and morphologically similar to the South Pacific Basin. This area may have acted as a nucleus for the eastward growth of the parallel island arcs between the Tasman Basin and the western edge of the South Pacific Basin at the circum-Pacific Belt.

Three guyots show relief of more than 14,000 feet. The depths of the guyot platforms range from 69 to 150 fathoms. These platforms are much shoaler than the 800-fathom platform depth of the guyots of the Western Pacific (Hamilton, 1956) or the 400- to 500-fathom platform depth of the guyots in the Gulf of Alaska (Menard and Dietz, 1951). The plat-

forms are probably associated with both eustatic sea-level change and subsidence of the guyots. The shoalest guyot, Taupo Bank, is thought to be Pleistocene, whereas the deeper guyots, Barcoo and Derwent Hunter, are thought to be mid- to late Tertiary. *Acropora*-type branching coral has been dredged from the Derwent Hunter Guyot at a depth of 196 fathoms. This guyot has subsided several hundred feet since the coral grew on its platform.

Elizabeth and Middleton reefs have grown on truncated volcanoes beveled by wave action during the Pleistocene glacial sea-level change. Capel Bank and Kelso Bank are table reefs probably formed in the same way except that coral growth was not fast enough to reach the surface as either the sea level rose or the sea mounts sank.

The coral reef along the west side of Lord Howe Island formed after the wave-cut platform surrounding the island was formed, and is Late Pleistocene to Recent.

The Recent coral-reef material along the southern part of the Great Barrier Reef is 530 feet thick; this plus the underlying Tertiary material totals 1795 feet.

The three guyots are aligned along a north-south line which probably represents a fault or shear zone in the thin ocean crust. The seamounts NT 1, NT 2, NT 3, are volcanic in origin, each having a total relief exceeding 11,000 feet, may lie along the same shear zone.

Lord Howe Island, Balls Pyramid, Elizabeth Reef, Middleton Reef, Capel Bank, Kelso Bank, and NT 5 all lie along a north-south line near the western edge, but separated from the main part, of Lord Howe Rise. These volcanic island, reefs, and seamounts are probably related to linear fracturing of the crust.

Lord Howe Rise has a 20-km crust and was probably formed by orogenic activity on an area of thin ocean-type crust. Lord Howe Rise is considered to have formed in the Early Paleozoic.

The Coral Sea Platform has a crustal thickness of 15–20 km and is more closely related to Lord Howe Rise than to the Tasman Basin.

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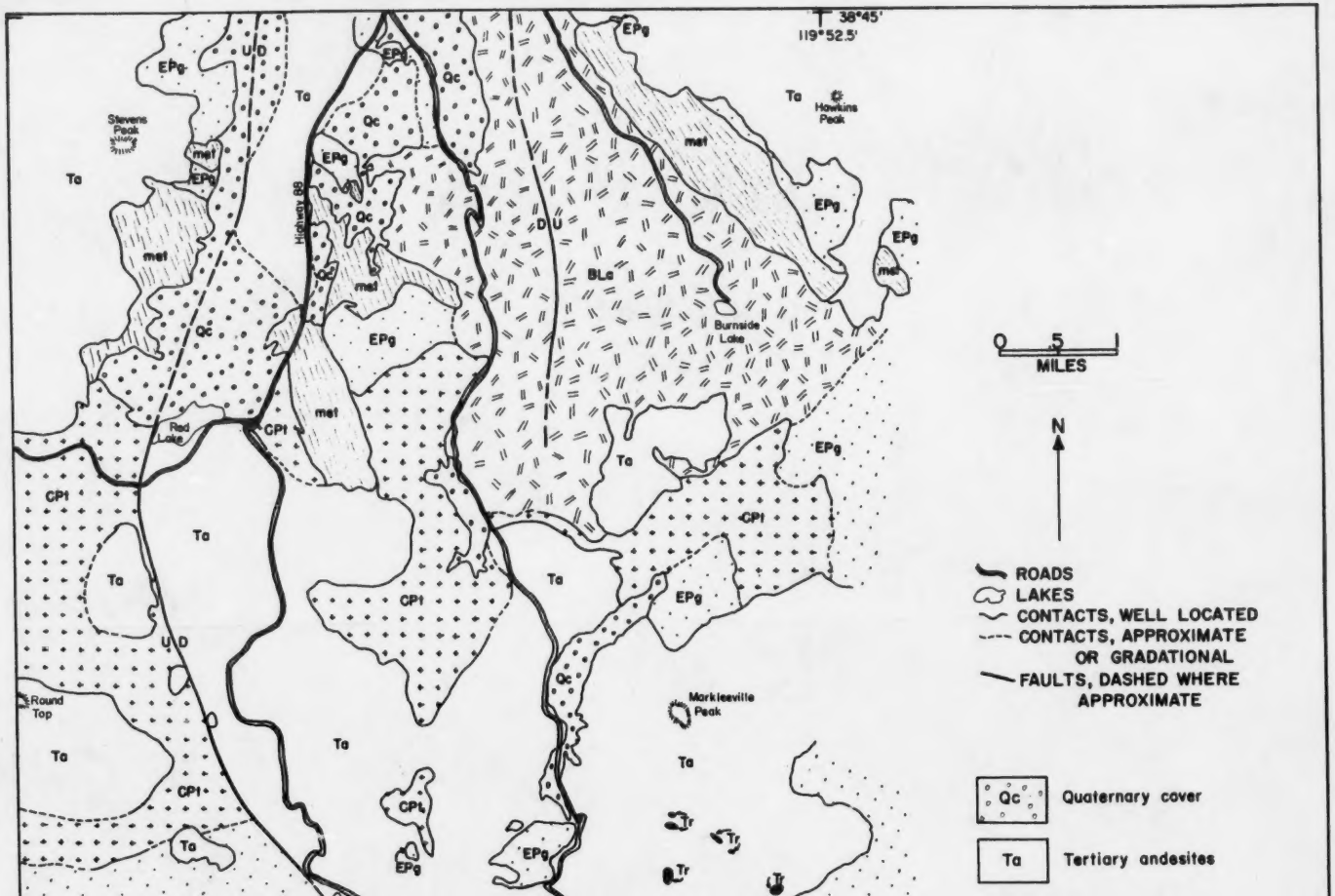
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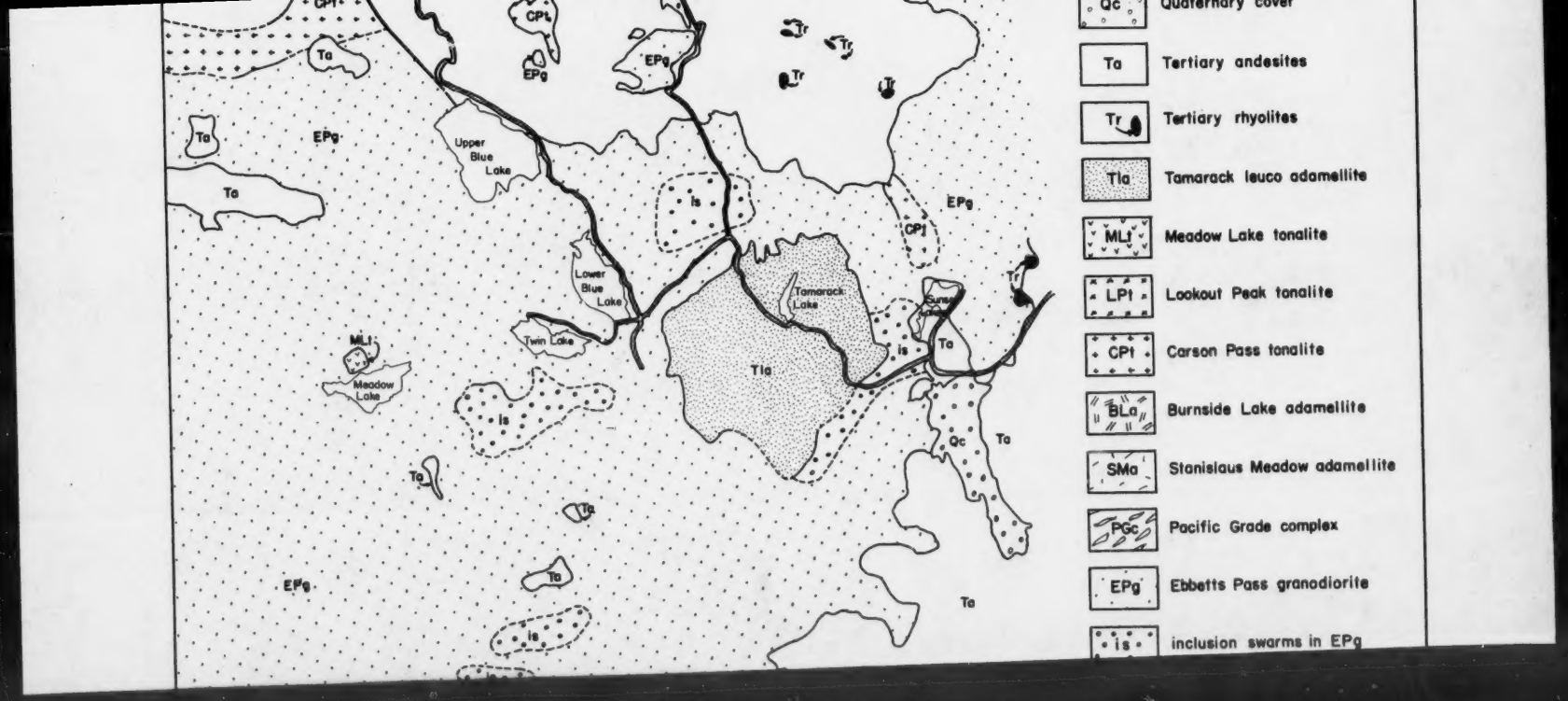
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Qc Quaternary cover
Ta Tertiary andesites
Tr Tertiary rhyolites
Tla Tamarack leuco adamellite
MLt Meadow Lake tonalite
LPt Lookout Peak tonalite
CPt Carson Pass tonalite
BLa Burnside Lake adamellite
SMa Stanislaus Meadow adamellite
PGc Pacific Grade complex
EPg Ebbetts Pass granodiorite
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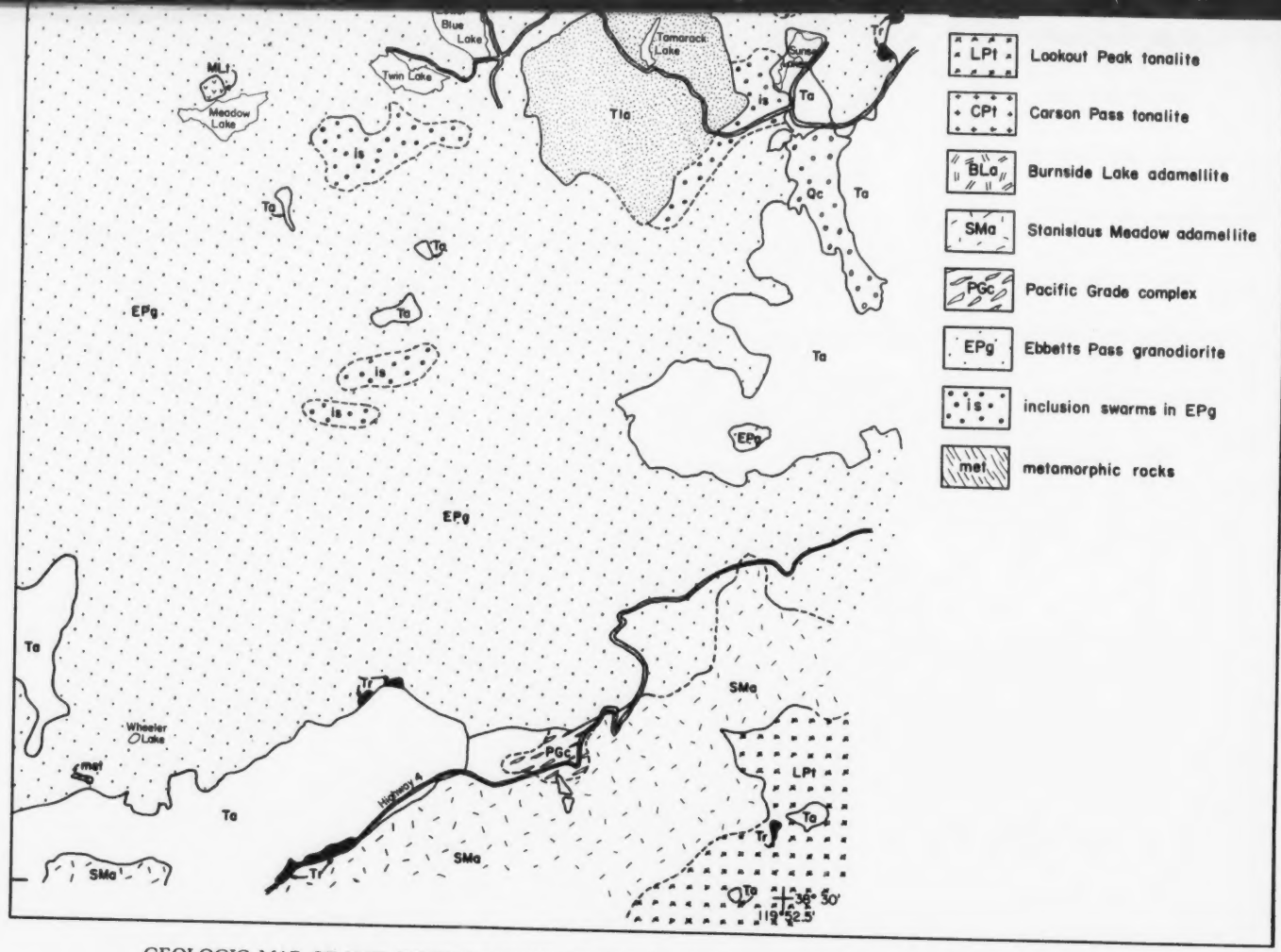

GEOLOGIC MAP OF THE WESTERN HALF OF THE MARKLEEVILLE (15′) QUADRANGLE, CALIFORNIA

PARKER, PLATE 1
Geological Society of America Bulletin, volume 72

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Petrology and Structural Geometry of Pre-Granitic Rocks in the Sierra Nevada, Alpine County, California

Abstract: The petrology and structural geometry of pre-granitic rocks in western Alpine County, California, were investigated by field and laboratory study. Pre-Cretaceous (Triassic?) metamorphosed equivalents of siltstones, sandstones, tuffs, limestones, andesite flows, and conglomerates are preserved in three roof pendants surrounded by a variety of Cretaceous (?) granitic rocks, principally adamellite and granodiorite.

The metamorphic rocks exhibit asymmetrical folds in the bedding (S_1), the axes (βS_1) of which trend northwest with an average plunge of 25°, and the axial surfaces of which dip steeply southwest. A statistically vertical tectonic surface (S_2) trends northwest and varies about axes (βS_2) approximately parallel to βS_1 in some fields, and inclined to βS_1 in others. Two sets of lineations are recognized, one parallel to βS_1 (B) and another plunging steeply at approximately right angles to βS_1 (B'). The geometric symmetry of both field and microscopic structures is triclinic. Deformation was probably by a combination of early flexural slip folding of S_1 about an axis parallel to βS_1, and later passive folding by slip on an intragranular scale along surfaces parallel to S_2. The slip on S_2 is thought to have been normal to the axis βS_2 in the plane of S_2. Strain of the mass involved shortening normal to the trend of the Sierra Nevada, accompanied later by extension in S_2. A regional study suggests that the orientation and style of the structural geometry of deformed rocks are similar in many parts of the Sierra Nevada, particularly near the present crest of the range.

Post-kinematic thermal metamorphism caused recrystallization of most of the tectonites into hornfelses, some of which show relict textures. All the metamorphic rocks are in the hornblende hornfels facies. Three metamorphic assemblages are recognized which are characterized by (1) the pair calcite-amphibole, (2) clinopyroxene, and (3) wollastonite. The three assemblages are thought to be present because of gradients in composition and partial pressure of fluid phases during metamorphism. Metasomatically introduced boron, tungsten, molybdenum, iron, copper, tin, and sulfur are locally important components.

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INTRODUCTION

The area studied, in western Alpine County, California, is almost entirely within the west half of the Markleeville (15′) quadrangle (Fig. 1) and ranges from about 7000 to 10,400 feet above sea level, including the crest of the Sierra Nevada at this latitude. It is accessible by Highways 4 and 88 except during winter months.

This investigation represents part of a study of pre-Tertiary rocks in the western half of the Markleeville quadrangle (R. B. Parker, 1959, Ph.D. thesis, Univ. Calif., Berkeley, p. 1–172). Regional mapping was based partly on aerial photographs and partly on a topographic map. Pre-granitic rocks were mapped in more detail on enlargements of the aerial photographs and topographic map, and by plane table.

This study had three major objectives: (1) to determine the mineralogy of the metamorphic rocks in order to establish their metamorphic facies, and to evaluate physical and chemical gradients present during the metamorphism; (2) to study the geometry of surfaces, lineations, and grain fabrics in the metamorphic rocks as a guide to the deformational history; and (3) to compare structural relationships with other parts of the Sierra Nevada.

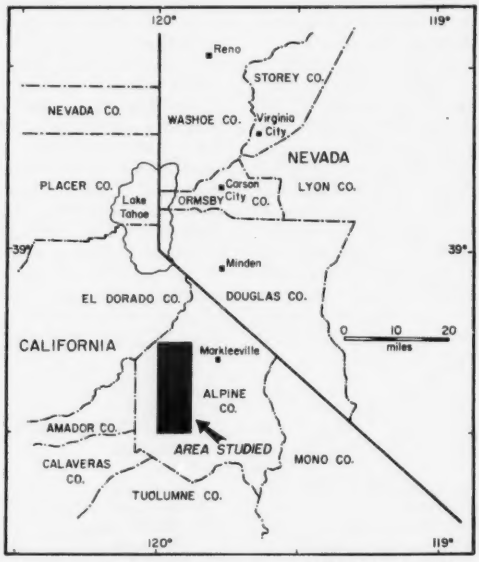

Figure 1. Index map. The western half of the Markleeville (15′) quadrangle is shaded.

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Curtis, F. J. Turner, L. E. Weiss, and Adolf Pabst of the University of California, Berkeley, for their help and encouragement. Field and laboratory studies were supported by the University of California and the National Science Foundation Predoctoral Fellowship program. The writer thanks Drs. D. L. Blackstone and R. S. Houston of the University of Wyoming, Dr. A. K. Baird of Pomona College, and Dr. F. A. Donath of Columbia University who read the manuscript and made numerous valuable suggestions.

GEOLOGIC SETTING

General Geology of the Sierra Nevada

The rocks of the Sierra Nevada were divided initially into two groups in the Mother Lode region (H. W. Turner, 1898; F. L. Ransome, 1900; A. Knopf, 1929, p. 8–22)—the bedrock series including metamorphic and granitic rocks, and the superjacent series including later volcanic and sedimentary rocks.

The bedrock series is composed of a western belt of metamorphic rocks with scattered granitic plutons and an eastern belt of granitic rocks containing metamorphic remnants.

The metamorphic rocks are divided (Taliaferro, 1943, p. 280–285; S. G. Clark, 1930, Ph.D. Thesis, Univ. Calif., Berkeley, p. 12–22) into a pre-Mesozoic formation, the Calaveras; and three Mesozoic formations, the Milton, Amador, and Mariposa. The rocks of these formations are similar and consist of slates, phyllites, metasandstones, marbles, cherts, and metavolcanic rocks. The degree of recrystallization of the Calaveras Formation is somewhat greater than that of the Mesozoic rocks.

The granitic rocks have been divided into two groups on the basis of potassium-argon dating (Curtis *et al.*, 1958, p. 5–9). The first group, 133–143 million years old, is mostly in the foothills (western) belt. The second group, 77-95 million years old, makes up most of the main batholith of the Sierra Nevada. The older rocks are principally tonalites and granodiorites; the younger rocks are principally granodiorites and adamellites.

The Tertiary superjacent series includes a variety of volcanic rocks (Eocene rhyolitic rocks and Mio-Pliocene andesitic rocks being most abundant), auriferous gravels, and related sediments. The rocks of the superjacent series occupy Tertiary stream channels and cover large areas of a Tertiary erosion surface. The reader is referred to Lindgren (1911), and Axelrod (1957).

Structure of the Bedrock Series

The structure of the bedrock series in the foothills belt was generalized (Taliaferro, 1943; unpub. ms.) as a series of isoclinal folds with axes trending parallel to the long dimension of the range and plunging at low angles with steep northeast-dipping axial planes. The attitudes of steeply dipping faults approximately parallel the axial planes of the folds. This generalization is valid regionally but there are many small-scale complexities (R. J. P. Lyon, 1954, Ph.D. thesis, Univ. Calif., Berkeley, p. 1–97; A. K. Baird, 1960, Ph.D. thesis, Univ. Calif., Berkeley, p. 1–102).

Taliaferro (1943, p. 285) and Compton (1955, p. 11) state that the plutons in the foothills belt both crosscut and conform to the metamorphic structures, and that some show relations suggesting the crowding aside of metamorphic rocks during emplacement. The complexity described in the Yosemite area (Calkins, 1930) is typical of many parts of the main batholith.

Pre-Tertiary History of the Sierra Nevada

The pre-Tertiary history of the Sierra Nevada has been only slightly refined from that proposed by Becker and his coworkers (Turner, 1898). Lack of reliable paleontologic dates has been partially overcome by radiometric dating (Curtis *et al.*, 1958). A summary of the pre-Tertiary history has been given by Compton (1955, p. 12–13) and Curtis *et al.* (1958, p. 10–13).

(1) late Paleozoic sedimentation and volcanism in a geosyncline approximately coextensive with the present Sierra Nevada

(2) probable Permian or early Triassic orogeny, with intrusion of basic and ultrabasic rocks, and perhaps some granitic rocks

(3) erosion and renewed marine deposition during late Triassic and up to late Jurassic

(4) late Jurassic (Nevadan) orogeny with synkinematic intrusion of ultrabasic rocks, and late synkinematic to postkinematic intrusion of granitic rocks of the older group

(5) upper Cretaceous (Santa Lucian) postkinematic intrusion of granitic rocks of the younger group.

Summary of the Pre-Tertiary Geology of Western Alpine County

The bedrock series is represented by both metamorphic and granitic rocks. The metamorphic rocks are the deformed and metamorphosed equivalents of Mesozoic sedimen-

tary and volcanic rocks, cropping out as three remnants of about 2 square miles each in surface exposure, and two much smaller areas. The three large masses are probably roof pendants and are surrounded by a granitic complex of intermediate to felsic rocks, within which eight types were mapped (Pl. 1).

PETROLOGY OF THE PRE-GRANITIC ROCKS

General Description and Age

Most of the rocks in the metamorphic bodies are derivatives of siltstones (calcareous in part), fine-grained feldspathic sandstones, marly tuffs, andesitic flows, and impure limestones; and, to a lesser extent, conglomerates, quartzose sandstones, and pure limestones. Primary stratigraphic relationships among the various mappable units are difficult to establish because of the intensity of subsequent deformation.

Most of the rocks are typical hornfelses, ranging from completely recrystallized granoblastic rocks to those that show relict clastic, volcanic, and tectonite textures. Most are uniformly even-grained, but a small proportion is porphyroblastic; schists are rare. Micas in some of the hornfelses show a preferred orientation and are locally abundant enough to impart a fissility to the rock.

No fossils were found, hence determination of age depends upon lithologic and structural correlations with fossiliferous rocks to the northwest and east. S. G. Clark (1930, Ph.D. thesis, Univ. Calif., Berkeley, p. 22–23) described very similar rocks to the northwest which he assigned to the Milton Formation of early Mesozoic age. Curtis (personal communication, 1959) found Triassic fossils in rocks of very similar lithology in the Pine Nut Range, Nevada. Since these rocks are intruded by granitic rocks of probable Cretaceous age, they are pre-Cretaceous, probably Triassic.

As there is considerable variation within map units, the rocks have been grouped into four chemical categories—calcareous, quartzo-feldspathic, basic (mafic) rocks as defined by Turner (Fyfe *et al.*, 1958, p. 200) and metasomatic rocks.

Calcareous rocks. The calcareous rocks are granoblastic aggregates of calcite and lime silicates, with and without free quartz. Garnet porphyroblasts, thought to be in the composition range $Gr_{92}Ad_8$ to $Gr_{74}Ad_{26}$ as indicated by refractive index and unit-cell dimensions (R. B. Parker, 1959, Ph.D. thesis, Univ. Calif., Berkeley, p. 20–21), accompanied by brown idocrase (N_o = 1.730–1.744) in quartz-free rocks, are set in a matrix of clinopyroxene (colorless, $2V_x$ = 52°–58°, N_y = approx. 1.68) calcic andesine, quartz, calcite, and wollastonite, with locally abundant microcline, graphite, and tremolite (colorless, $2V_x$ = 84°–89°). Scapolite porphyroblasts (N_o = 1.564–1.574) are locally very plentiful in both calcite-rich and feldspathic rocks. Sphene (reddish, pleochroic) and epidote (green, $2V_x$ = 84°–89°) are common minor accessories. Green spinel and prehnite are present locally but are nowhere abundant.

Quartzo-feldspathic rocks. The quartzo-feldspathic rocks embrace the bulk of the derivatives of siltstones and sandstones. They are fine-grained (0.01–0.5 mm) granoblastic rocks composed of quartz, oligoclase, and microcline, with small amounts of clinopyroxene and tremolite-actinolite amphiboles. Micas are relatively abundant (<10 per cent) in about one-fourth of the specimens examined; and sphene, graphite, and pyrite (or alteration products thereof) are common minor accessories. Cassiterite, topaz, and copiapite are locally present in trace amounts.

Mafic rocks. The mafic rocks are derivatives of massive volcanic rocks or tuffs. The distinction depends on texture, as the mineralogy of both types is similar. Major minerals are plagioclase (andesine to labradorite), amphibole ($2V_x$ = large, strongly pleochroic in shades of green and brown), and biotite ($2V_x$ = small, r>v, pleochroic colorless to red brown or brown). Chlorite ($2V_x$ = small, pale green, anomalous interference colors in shades of brown and purple) is closely associated with biotite in some specimens. Pyrite and brown sphene are common minor accessories.

Metasomatic rocks. The metasomatic rocks are of two types—one with evidence of the introduction of boron, and the other of tungsten, molybdenum, copper, and iron.

Many of the rocks in the central roof pendant are very rich in boron-bearing minerals. Axinite makes up more than 90 per cent of many calcareous rocks, and tourmaline is equally abundant in many mafic and quartzo-feldspathic types. The distribution of the boron-rich rocks is not systematically related to visible granitic contacts.

In contrast, the tungsten, molybdenum, copper, and iron-rich rocks are restricted to the immediate vicinity of metamorphic-granite contacts. They are mostly typical skarns com-

posed of iron-rich garnet, green clinopyroxene, and strongly pleochroic amphibole, together with quartz, calcic plagioclase, and chlorite. Magnetite, pyrrhotite, chalcopyrite, scheelite, pyrite, and molybdenite are locally plentiful.

Metamorphic Facies and Zonation

The metamorphic rocks are placed in the hornblende hornfels facies of contact metamorphism (Fyfe *et al.*, 1958, p. 206–211). The assemblages containing micas eliminate the pyroxene hornfels facies, and those containing Ca-bearing plagioclase and clinopyroxene eliminate the albite-epidote hornfels facies.

A broad zonation of the metamorphic assemblages may be defined as follows: *zone 1* rocks are characterized by the occurrence together of calcite and amphibole without clinopyroxene; scapolite is very abundant; *zone 2* rocks are characterized by the occurrence of clinopyroxene; *zone 3* rocks are characterized by the occurrence of wollastonite (Fig. 2).

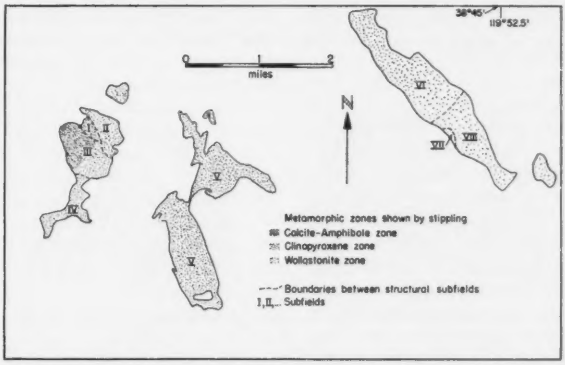


Figure 2. Map of metamorphic bodies showing distribution of metamorphic zones and structural subfields. The zones are characterized by the occurrence of (1) the pair calcite-amphibole, (2) clinopyroxene, and (3) wollastonite.

Inferences Concerning Conditions of Metamorphism

Speculations concerning the conditions of metamorphism can be based on stability fields of some of the minerals. The presence of scapolite and tremolite (Fyfe *et al.*, 1958, p. 160–162) and wollastonite (Danielsson, 1950, p. 63; Harker and Tuttle, 1956, p. 249) suggest a temperature of metamorphism between 400°C and 700°C, assuming a load pressure of about 1000 bars. Whereas, in view of the small sizes of the metamorphic masses, it seems unlikely that large gradients of either temperature or load pressure could have been present during metamorphism, the fluid pressure may have varied, and the fluid phase may have been of different composition in different parts of the masses. The abundance of scapolite in zone 1 suggests a relatively high partial pressure of halogens and carbon dioxide. The presence of hydrous minerals in most assemblages may indicate that the partial pressure of water was appreciable throughout or that the temperature was low. The presence of wollastonite in zone 3 suggests a relatively low partial pressure of carbon dioxide locally, possibly related to a relatively high permeability of the rocks during metamorphism.

STRUCTURAL GEOMETRY OF THE PRE-GRANITIC ROCKS

Methods and Terminology

The method of study of the structural geometry of the metamorphic rocks combined two techniques: (1), mapping the distribution of rock units and direct observation of major folds in the field; and (2) statistical study of structural elements measured in the field. The relative importance of the two techniques is different in the various areas studied because of differences in degree of exposure. The geologic map of the western roof pendant (Pl. 2) is

typical of the three metamorphic masses. Some of the attitudes measured are mapped.

Terminology, essentially that of Bruno Sander, is discussed (in English) by Knopf and Ingerson (1938, p. 1–85), Turner (Turner and Verhoogen, 1960, p. 602–654), and Weiss (1959a, p. 92–93). Penetrative surfaces are called S (Turner and Verhoogen, 1960, p. 620–621), with subscripts to distinguish the two types recognized. Small folds and other lineations are collectively called L and may be divided into two groups—B with shallow plunges, and B' with steep plunges. The letter β is used for a statistically defined axis of intersection of a group of surfaces and has the significance of a fold axis in some cases. The axis β possesses only orientation, and not position. All projections of structural data are lower hemisphere, equal-area, and were prepared by use of a 20 cm printed net.

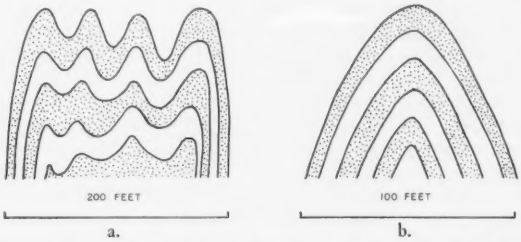


Figure 3. Diagrammatic sketches of two styles of folds. a. Isoclinal fold with broad hinge zone characterized by numerous minor folds and undulations; b. Closely appressed fold with simple hinge

Description of Structural Elements

Bedding. Bedding (S_1) is locally well preserved. Criteria that suggest that lithologic layering is bedding are: (1) continuity of lithology in the plane of the layer; (2) presence of sedimentary structures, especially graded bedding.

Bedding is in general well preserved in derivatives of impure limestones, lava flows, conglomerates, and massive sandstones with shale interbeds and is mainly absent or highly distorted in derivatives of tuffs, pure limestones, and siltstones.

Foliation. The term foliation is used here to designate only secondary foliation and, more specifically, any tectonically produced surface. The foliation (S_2) is of two types, depending on the rock type.

(1) Thin lithologic layering (<2 cm), probably transposed bedding (primary lithologic layering that has been rotated to parallelism with slip surfaces: Knopf and Ingerson, 1938, p. 189–190), is present in metasiltstones, marbles, and some of the calc-silicate rocks. It is characterized in these rocks by parallel, discontinuous lenticles of contrasting rock types which are disposed both aligned and en echelon.

(2) Fissility or cleavage is produced by a preferred orientation of platy minerals in a few metatuffs and metasiltstones.

In most places the foliation is a surface of slip. Small offsets of bedding layers that transgress the foliation are widespread. The sense of slip on the foliation is not constant, and small passive[1] folds in the bedding are present locally.

Folds. Two styles (Weiss, 1954, p. 11–12) of folds in the bedding are recognized, not including small-scale folds which are treated as lineations:

(1) Isoclinal folds with broad complex hinges made up of many small folds and undulations are found in calc-silicate rocks, in the eastern pendant, and in metavolcanic rocks, in the western pendant. Beds in the calc-silicate rocks are thinner on the limbs than on the hinge (Fig. 3a).

(2) Closely appressed folds with simple hinges are found in less massively bedded calc-silicate marbles in the western pendant (Fig. 3b).

Very ductile rocks such as marbles and some of the metasiltstones show no folds since bed-

[1] A passive fold is one in which the folded layer behaved as a passive indicator layer during the folding (the shear or slip fold of Turner and Verhoogen, 1960, p. 607).

ding is almost obliterated. Pure marbles show abundant evidence of flow such as "intruded" satellitic dikes into adjacent units and flow of marble into spaces between boudins of siliceous marble.

The orientation of βS_1 (the mean axis of intersection of a number of attitudes of S_1 in projection, the equivalent of the pole to a π girdle) was determined graphically for eight subfields (a subfield is a three-dimensional subdivision with approximate structural homogeneity) by plotting the attitudes of S_1 on a lower hemisphere, equal-area net, and contouring the points resulting from the intersections of the projected planes. Lineations measured in the same subfield are plotted as points on the βS_1 diagrams (Fig. 4). The significance of the weak maxima in some of the βS_1 diagrams may be indicated by comparing the orientation of the βS_1 axes in subfields VI and VIII with that in subfield VII. Subfield VII is a fold in calc-silicate marbles in which a variety of attitudes on all parts of the fold defines a βS_1 axis of superior reliability which agrees with the orientation of the fold axis estimated in the field by observation. Subfields VI and VIII, in the same roof pendant, contain rocks in which S_1 is nearly parallel throughout, and the maxima in the βS_1 diagrams are weak. Nevertheless, the βS_1 axes defined in subfields VI and VIII are almost identical in orientation with that in subfield VII. From this agreement in orientation (approx. N. 32° W. 20° NW.) it is concluded that the βS_1 axes have the significance of fold axes in the sense of Clark and McIntyre (1951, p. 594), according to whom "The axis of a fold is defined as the nearest approximation to the line, which, moved parallel to itself in space, generates the fold."

The βS_2 diagrams, prepared in the same way as the βS_1 diagrams, may be divided into two groups,—those in which the βS_2 axis approximately parallels the βS_1 axis for the same subfield (subfields I, III, VI, VII, and VIII), and those in which the βS_2 axis is inclined at a large angle to the βS_1 axis for the same subfield (subfields II, IV, and V). In the latter group the βS_2 axes are steeply plunging.

Lineations. The term lineation here includes a number of linear structures (collectively called L), visible on the scale of an outcrop, divided into five types:

(1) Small-scale corrugations in fields of distorted or transposed bedding appear to be irregularities in the surface of breaking of the outcrop caused by intersections of separate S_2 surfaces having slightly different orientations about a common axis (Fig. 5a).

(2) Mineral tracts or trains made up of aggregates of mica, garnet, or green spinel, appear to be smeared-out or elongated aggregates, but may be due to intersections of transposed bedding with S_2 as described from other Sierran areas by R. J. P. Lyon (1954, Ph.D. thesis, Univ. Calif., Berkeley, p. 21) and M. L. Christensen (1959, Ph.D. thesis, Univ. Calif., Berkeley, p. 58–59) (Fig. 5b).

(3) S_1/S_2 intersections that look like fine striping on many surfaces and resemble the corrugations where bedding is offset on S_2 (Fig. 5c).

(4) Boudins and boudin-like structures are common, but are not measureable in most places (Fig. 6). The unmeasureable boudins seem to have their long axes parallel to βS_1. In a marble body northwest of Crater Lake (Pl. 2) boudins are formed from siliceous layers in the otherwise calcite-rich marble. In addition to true boudins there is a pinch-and-swell structure in many layers in the eastern pendant which defines a crude lineation approximately parallel to βS_1. A third boudin-like structure in the eastern pendant is a very irregular pinch-and-swell of lithologic layers resembling the artificial structures produced by Ramberg (1955, p. 521, Pl. 6c).

(5) Small folds, locally present where bedding is well preserved, show characteristics of both flexural slip folding and passive folding.

These lineations are found mostly in one of two orientations, although a number are intermediate. Some are approximately parallel to βS_1 (called *B* lineations), and some are steeply plunging at high angles to βS_1 (called *B′* lineations) (Fig. 7).

Joints. Attitudes of joints were measured in the eastern pendant (subfields VI, VII, and VIII) (Fig. 9a). Two types of joints are recognized; the distinction depends upon their orientation with respect to folds and lineations observed in the field: (1) those approximately normal to βS_1 and the *B* lineations, called *ac* joints; (2) those approximately normal to the *B′* lineations, called *a′c′* joints (Turner and Verhoogen, 1960, p. 623–624). Both types are commonly present in a single outcrop. Evidence as to the order of formation of the two sets, as judged from termination of one joint at another joint surface, is contradictory, and on the scale of the whole roof pendant, they appear to have developed contemporaneously. The joints are simple, unmineralized fractures with no evidence of movement parallel to

them. The development of the joints in these orientations may reflect the rock fabric; they were not necessarily formed during the main period of deformation.

Microfabric. The microfabrics of five specimens were examined by standard universal-stage methods to investigate the relation of grain fabrics to larger-scale structural elements. Most of the rocks are unsuited to petrofabric study because of their small grain size.

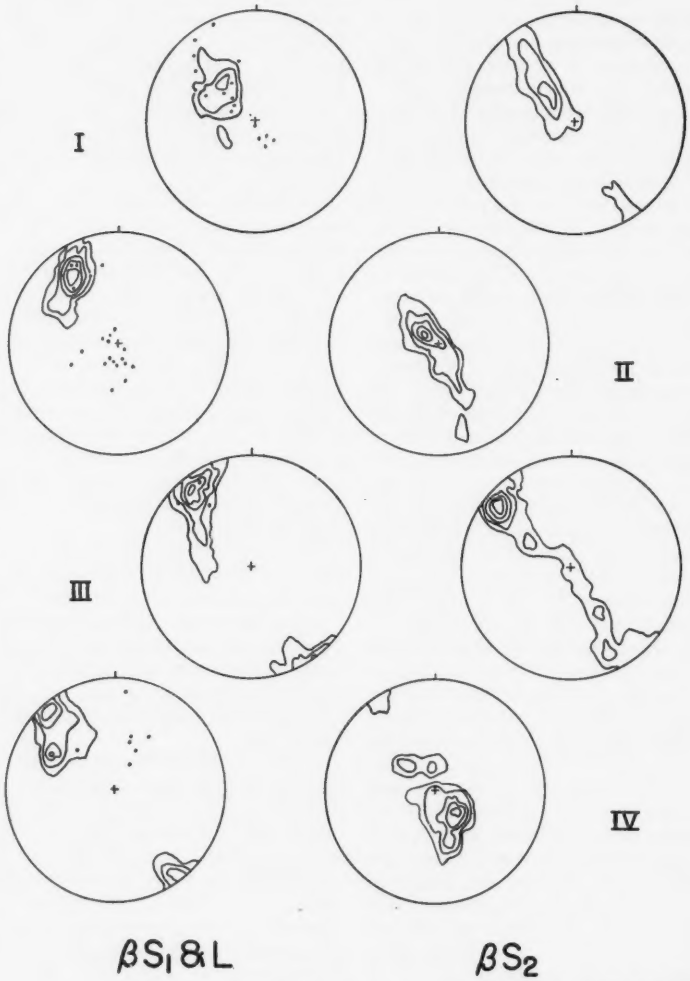
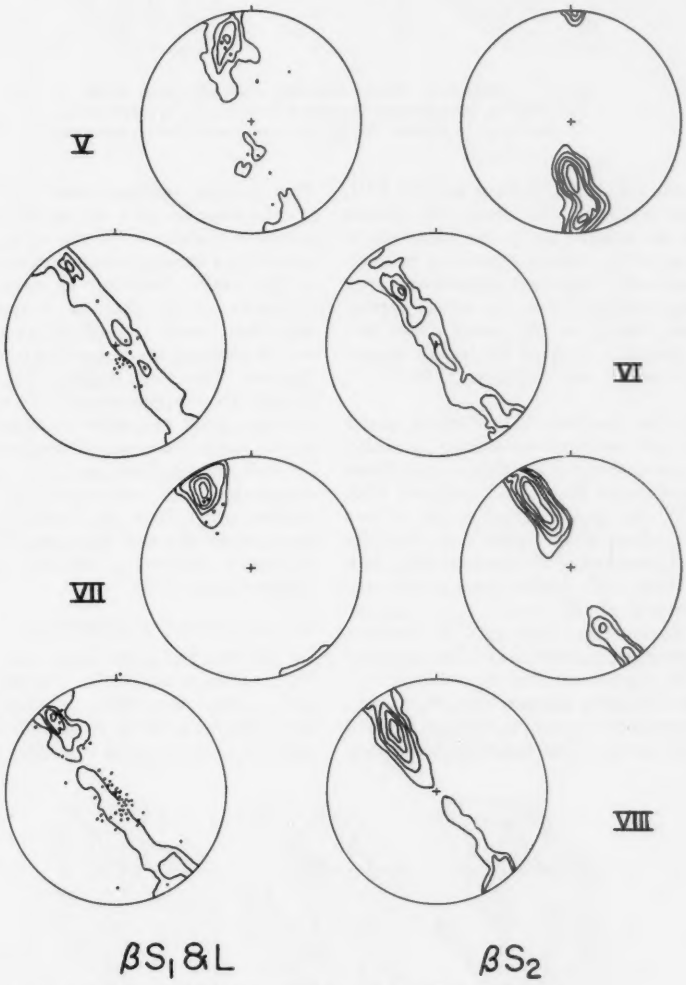


Figure 4. β diagrams showing the geometry of S_1, S_2, and L. Diagrams are contoured at 3–6–9–12–15 per cent per one per cent area. Tick marks show true

Subfield	S_1	S_2	L
I	49	43	15
II	44	25	23
III	42	17	3
IV	24	12	6

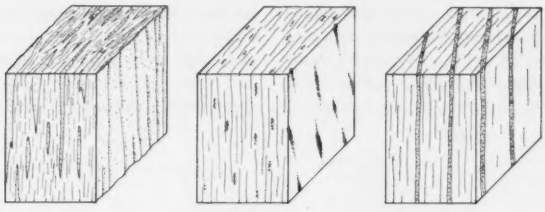
See Figure 2 for location of subfields.

The orientation of quartz *c*-axes was measured in five specimens, supplemented by poles to muscovite cleavage (Fig. 8c) and calcite *c*-axes (Fig. 8d) in specimen 405 (Univ. Calif., Berkeley specimen numbers) from subfield I in the western roof pendant. Specimen 405 is composed of equigranular calcite, minor (1–2 per cent) muscovite and clastic quartz grains, and traces of scapolite. Two thin sections at right angles were used. The other four speci-



north. All projections are lower hemisphere, equal-area plots. Number of data are as follows.

Subfield	S_1	S_2	L
V	42	8	10
VI	35	19	19
VII	40	9	6
VIII	26	18	46



a. b. c.

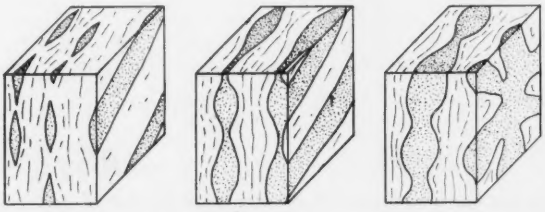
Figure 5. Schematic block diagrams showing three types of lineations. a. corrugations; b. mineral tracts; c. S_1/S_2 intersections as seen on an S_2 surface. Blocks are approximately 6 inches wide.

mens (166, 178, and 285 from subfield VIII; 284 from subfield VI), from the eastern pendant, are quartzo-feldspathic hornfelses in which part of the quartz appears to have recrystallized into its present granoblastic state from larger lensoid grains, the overall appearance being similar to the quartz grains in a Saxony granulite pictured by Sander (reproduced in Knopf and Ingerson, 1938, Pl. 3, fig. 2).

Except for specimen 405 in which quartz seems to lack preferred orientation (probably because it is surrounded by calcite as noted from other localities by Knopf and Ingerson, 1938, p. 64–65), the quartz diagrams are of two types: (1) those of specimens from outcrops showing a pronounced B' lineation (Fig. 8a is typical) show crude girdles about a steep axis approximately parallel to B'; (2) a diagram from a specimen not showing a B' lineation shows a weak girdle about an axis approximately parallel to the βS_1 axis for that subfield (VI) (Fig. 8b). The mica diagram (Fig. 8c) shows a strong maximum of poles to cleavage which is not at 90° to the visible foliation in the rock. Two possible interpretations are indicated; (1) the mica lies in a slip surface system not otherwise visible, or (2) the micas have been rotated to a limiting orientation by slip parallel to the visible foliation. In either case the symmetry of the diagram is approximately monoclinic with a girdle of poles about a steeply plunging axis, suggesting rotation about that axis. The calcite diagram (Fig. 8d) shows a girdle about approximately the same axis as the mica girdle axis, with a maximum normal to the visible foliation. Although this marble has been recrystallized, and is not completely comparable with experimentally deformed marbles which have not been recrystallized, the maxima of *c*-axes may parallel an axis of maximum shortening during deformation (Turner *et al.*, 1956).

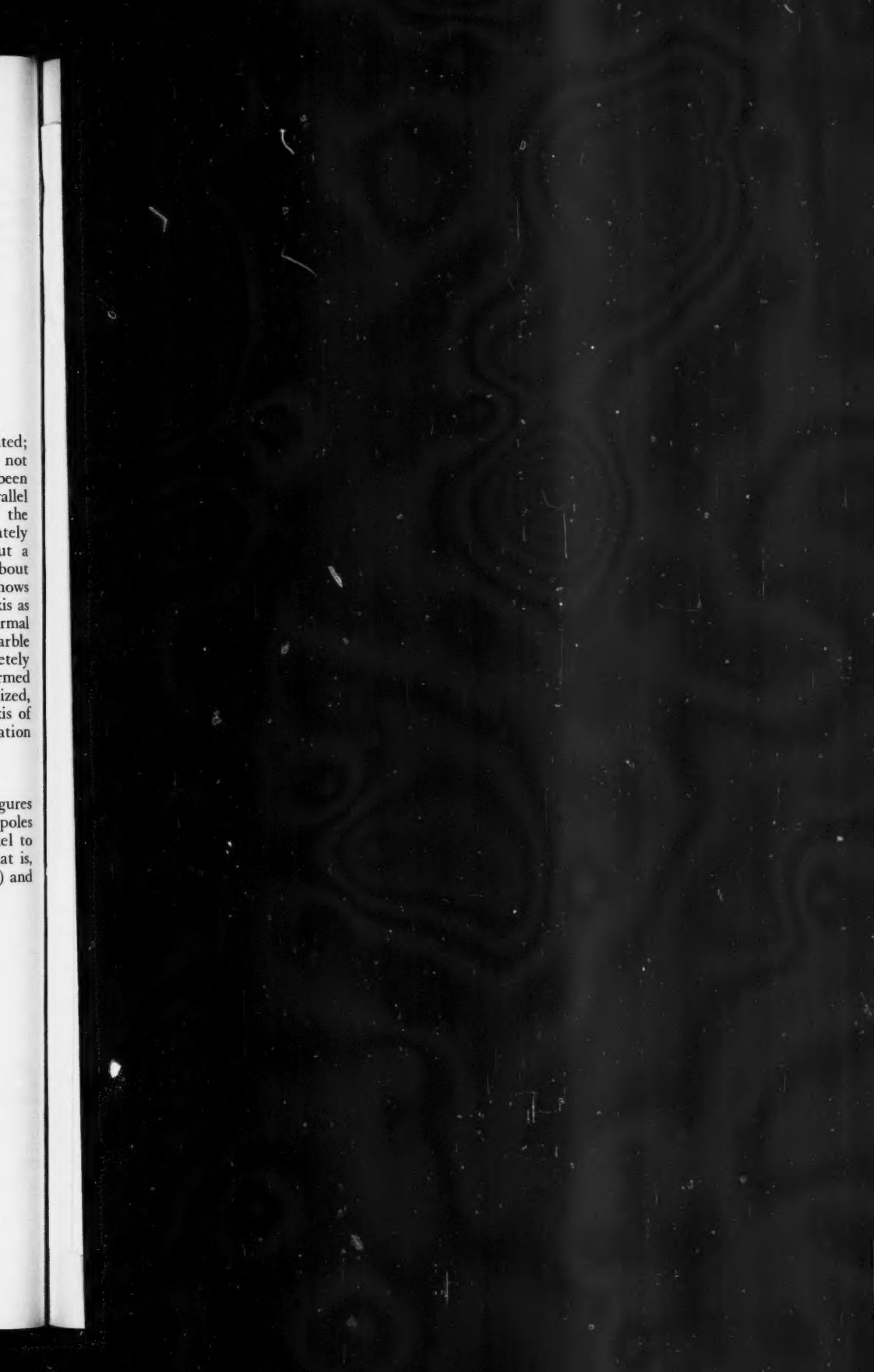
Synoptic Geometry and Symmetry

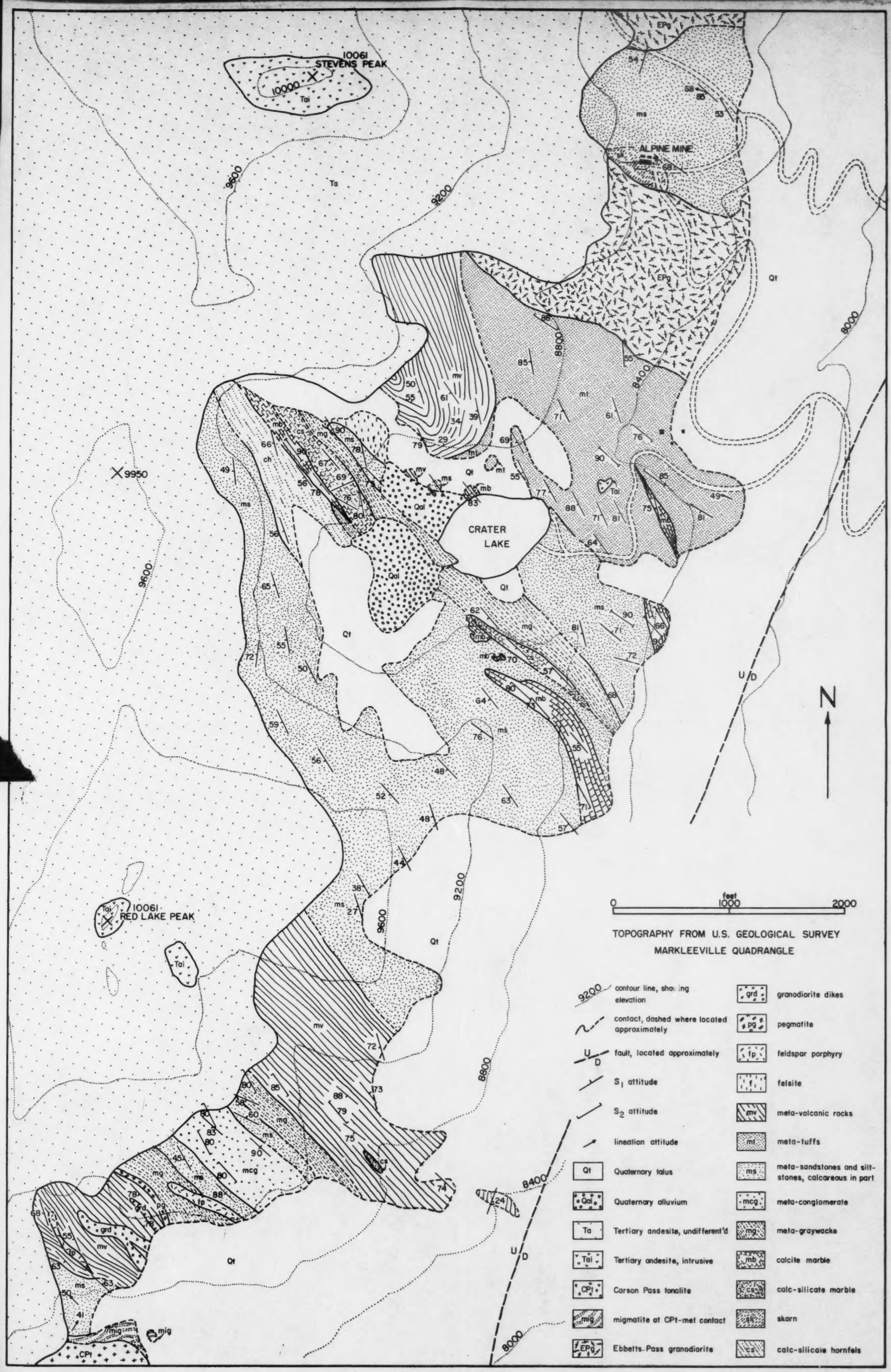
Collective S_1, S_2, and L are shown in figures 9b, c, and d, respectively. The girdle of poles to S_1 is consistent with a βS_1 axis parallel to those defined in the several subfields (that is, the pole of the π girdle is parallel to βS_1) and



a. b. c.

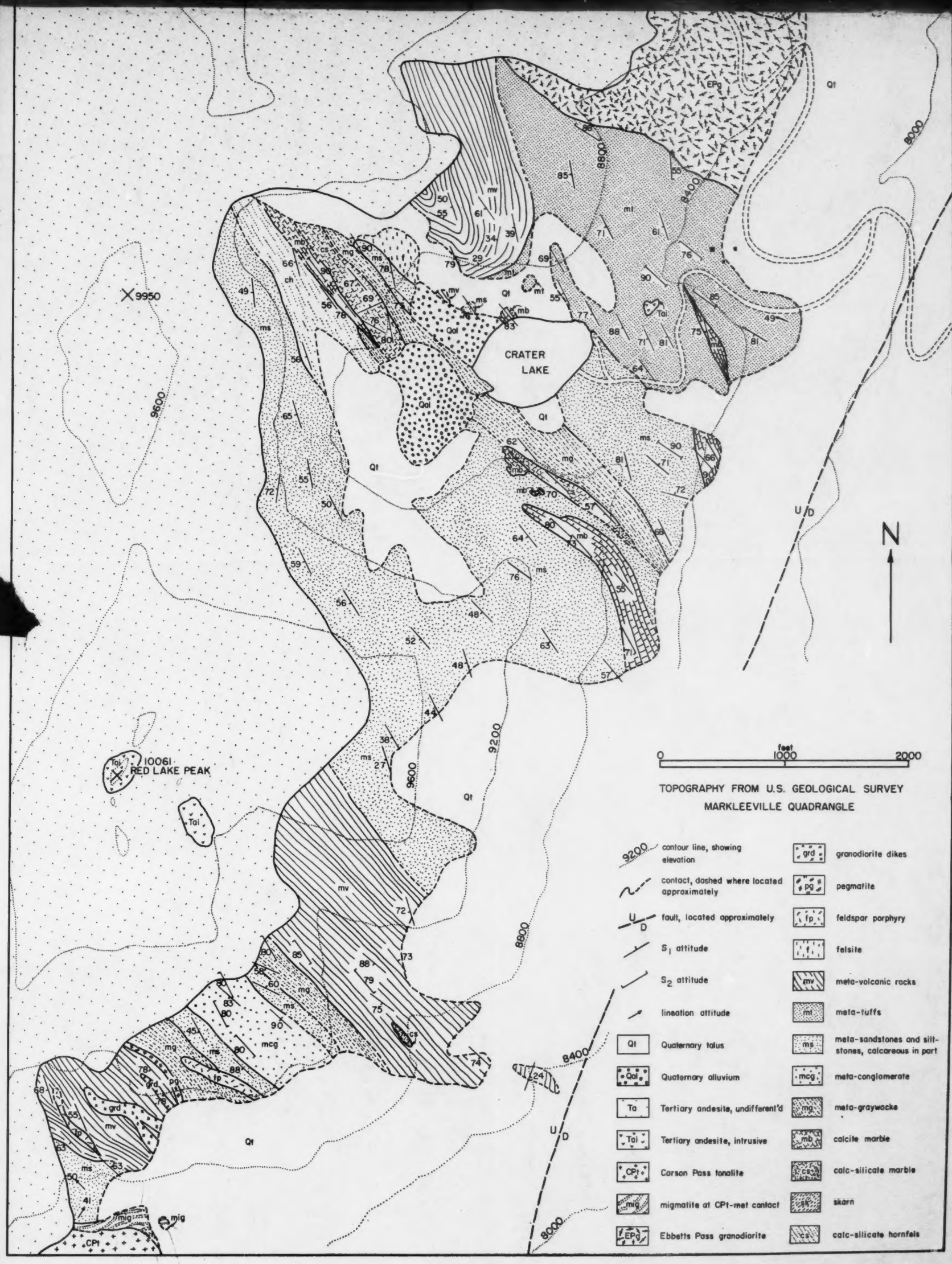
Figure 6. Schematic block diagrams showing three boudin-like structures. a. boudins developed from siliceous layers in marble; b. pinch-and-swell structures defining a lineation, as developed from sandstone layers in siltstone; c. irregular pinch-and-swell structures as developed in siltstones. Blocks are approximately 12 inches wide.





GEOLOGIC MAP OF THE WESTERN ROOF PENDANT

This map shows the type of outcrop pattern found in all three masses. Only a portion of the attitudes measured are mapped.



GEOLOGIC MAP OF THE WESTERN ROOF PENDANT

This map shows the type of outcrop pattern found in all three masses. Only a portion of the attitudes measured are mapped.

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shows a single maximum concentration which is not in the horizontal plane. The foliation, S_2 (Fig. 9c), is statistically vertical, with a slight spread in a girdle about a steep axis parallel to B', and strikes approximately parallel to the trend of the βS_1 axes or the pole to the πS_1 girdle (Fig. 9b). The lineations, L, are mostly of the pre-existing flexural folds. This relation is difficult to evaluate in the field but does not appear to be strongly supported by visible large folds.

In the second case (Fig. 10b), folds of S_1 have limbs of unequal length and are part of a larger anticlinorium culminating to the east.

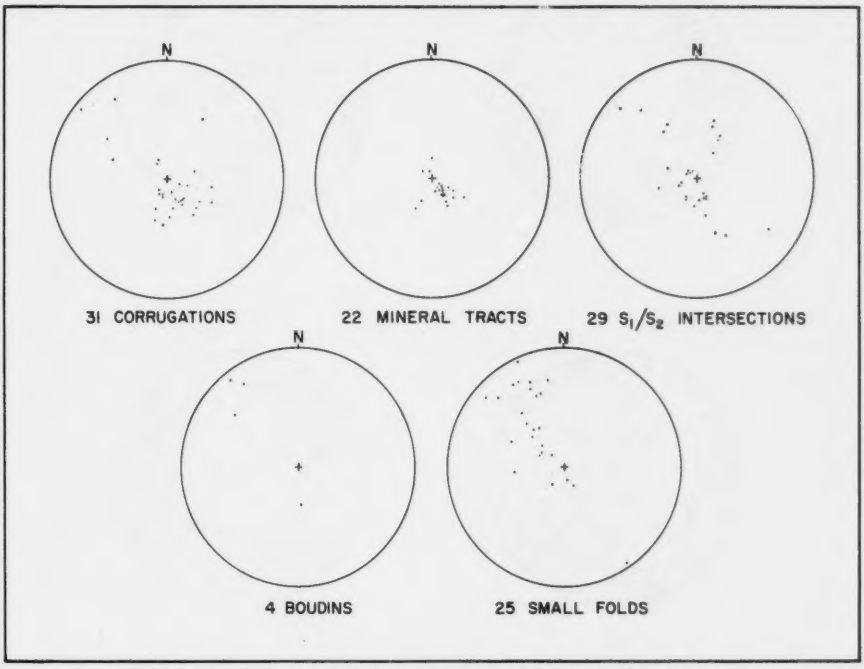

Figure 7. Projections showing orientations of five types of lineations

B' lineations, not parallel to the regional βS_1 axis, but tending to lie in S_2.

Two reasonable interpretations of the positions in space of S_1 (that is, the shape or style of the folds of S_1) are possible from the orientation data shown in Figures 9b and 9c (Fig. 10). In both cases it is assumed that S_2 is a vertical surface and that the folds initially developed as flexural slip folds and were later modified by slip on or parallel to S_2.

In the first case (Fig. 10a), folds in S_1 are asymmetrical, *after* passive folding, with axial planes dipping steeply southwest. Thus S_2 is not axial-plane foliation after passive folding although it might have approximated axial-plane orientation prior to passive modification

In contrast to the rather special nonaffine slip on S_2 required by the first case, the slip on S_2 would be statistically affine in this case with the greatest relative vertical transport or extension to the east. In this case S_2 would have remained axial-plane foliation throughout the passive deformation. This interpretation is strongly supported by the fact that map units can be mapped around tight hinges but do not reappear in other parts of the roof pendants. The metavolcanic rocks northwest of Crater Lake (Pl. 2) are a good example of this geometry. They are exposed in a large synclinal fold which appears to connect with a strongly appressed anticlinal structure to the west in an area of talus cover. However, such rocks are

not exposed either east or west of the paired syncline and anticline (the metavolcanic rocks in the southern part of the map are of different character). This geometry is well illustrated by Weiss (1959b, p. 20), who shows two possible fold forms that would give the same π diagram.

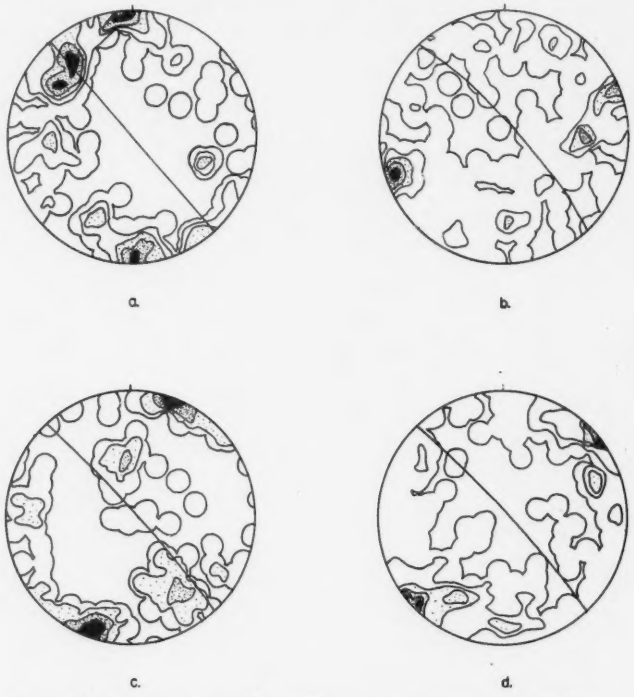

Figure 8. Equal-area plots of microfabrics. Megascopic surfaces (S_2) shown as great circles. Tick marks show true north. a. 151 quartz *c*-axes from specimen 166 contoured at .7–2–3–4–5 per cent per one per cent area; b. 200 quartz *c*-axes from specimen 285 contoured at .5–2–3–4–5 per cent per one per cent area; c. 106 poles to muscovite cleavage from specimen 405 contoured at .9–2–4–5 per cent per one per cent area; d. 175 calcite *c*-axes from specimen 405 contoured at .6–2–3–4–5–6 per cent per one per cent area

The geometry of the surfaces S_1 and S_2 and that of the lineations is triclinic if all are considered together, although either system of surfaces is approximately monoclinic if considered alone.

Interpretation

Movement plan. The symmetry of the movement plan may in many, if not all, cases be assumed to be reflected in the symmetry of the fabric (Turner, 1957, p. 2–4). Using this symmetry principle on a regional scale, the movement plan may be described as a rotation of S_1 about βS_1 (the *b* kinematic axis at this stage) in the early stages of the deformation history. The form of many folds suggests that this rotation was effected by flexural slip (Turner and Verhoogen, 1960, p. 643). Further folding of S_1 (the passive folding stage) by slip on S_2 was probably important after the development of S_2 as evidenced by the presence of small passive folds in many places. Slip was probably statistically affine.

Presumably S_2 would not have developed as

a surface of slip until S_1 had rotated into orientations unfavorable for continued slip parallel to bedding. The late movement of S_2 seems to have been active on an intragranular scale as well, as evidenced by the agreement of field with shallow βS_2 axes suggests that slip about a steep axis may have been important in all parts of the masses to some degree, probably in the late stages of deformation. It is further suggested that the steeply plunging regional

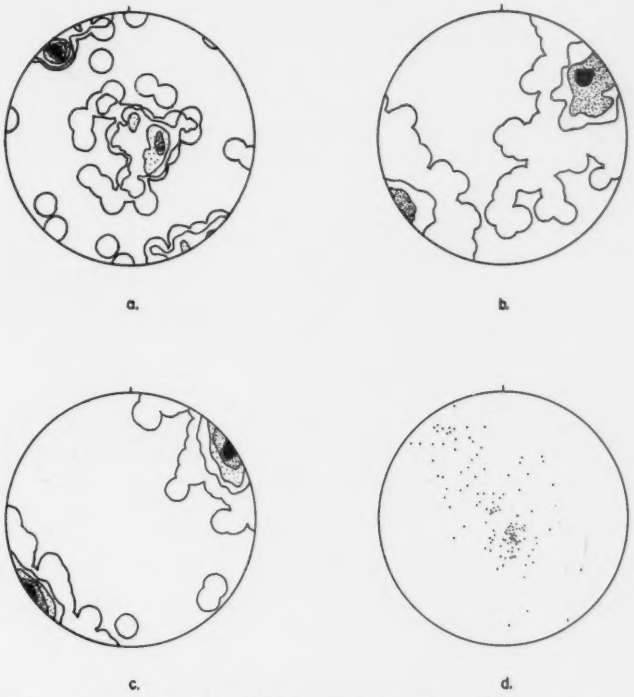

Figure 9. Equal-area plots of field data. Tick marks show true north. a. 63 poles to joints from the eastern roof pendant contoured at 1.6–3–6–9–12 per cent per one per cent area; b. 308 poles to S_1 contoured at .3–3–6–9 per cent per one per cent area. Diagram shows a maximum inclined to the horizontal in a girdle about an axis plunging northwest; c. 144 poles to S_2 contoured at .7–5–10–15–20 per cent per one per cent area. Diagram shows a horizontal maximum in a weak incomplete girdle about a steeply plunging axis; d. 128 lineations showing a concentration near the vertical and a spread in a girdle parallel to the plane of maximum concentration of S_2.

and microfabric data. The fact that S_2 defines β maxima instead of girdles indicates that slip on S_2 was not confined to a system of parallel planes but, rather, to a system of subparallel planes with a common axis, βS_2 (analogous to the pure shear concept), further supported by the corrugation lineations. The fact that steep linear elements are found even in subfields

βS_2 axis (parallel to the pole of the weak πS_2 girdle in Figure 9c) was the kinematic b axis during the passive folding stage of the deformation.

Strain. Shortening normal to the axial planes of the flexural slip folds was probably followed by flattening normal to the foliation as evidenced by the pinch-and-swell structures

and suggested by the calcite microfabric. Steeply plunging B' lineations may parallel a direction of preferred extension in S_2, indicated by the fact that the long axes of most of the boudins and boudin-like structures are normal to the B' lineations. This extension parallel to B' would have taken place during the passive folding stage and would thus be the familiar b extension as related to the symmetry of S_2.

Comparison With Other Parts of the Sierra Nevada

A regional study was made of the Sierra Nevada to establish whether the geometry of the Alpine County rocks is comparable with that of other Sierran tectonites. The data were drawn from published and unpublished sources, supplemented by brief examinations of a number of areas. References and brief descriptions of each of the 35 areas shown in Plate 3 are given in the appendix. All diagrams are β diagrams contoured 3–6–9–12–15 per cent per one per cent area. Lineations are shown as dots on the β diagrams. It should be emphasized that the diagrams represent the geometry of isolated fields, and that these fields are by no means a contiguous network.

The wide variation in orientation of β maxima shown in Plate 3 reveals the complexity of the geometry of surfaces in the metamorphic rocks of the Sierra Nevada. It is striking to note how commonly steep lineations are present, even in fields with a shallow β axis. In general, β axes with shallow or horizontal plunges are not common. The widespread presence of steep β axes and steep lineations suggests that most parts of the Sierra Nevada have been subjected to more than one deformation or to a single complex deformation such as suggested for the Alpine County rocks. The diagrams from fields in the high parts of the range are especially similar to the βS_2 diagrams in the Markleeville quadrangle.

SUMMARY

A layered succession of Triassic(?) sedimentary and volcanic rocks was intensely deformed to a series of tightly folded tectonites. The deformation was complex and seems to have been a combination of early flexural folding followed by deformation by slip parallel to a system of subparallel penetrative tectonic surfaces sharing a common axis. Evidence suggests that the rocks were on the southwest flank of a major anticlinorium. The complex structural pattern in Alpine County is very similar to the structural patterns in many other areas of pre-Cretaceous rocks in the Sierra Nevada.

Post-kinematic thermal metamorphism caused recrystallization of the tectonites into hornfelses. The mineral assemblages in the hornfelses suggest a temperature of metamorphism between 400°C and 700°C and are typical of the hornblende hornfels facies. Distribution of characteristic minerals indicates gradients in composition of a fluid phase during metamorphism.

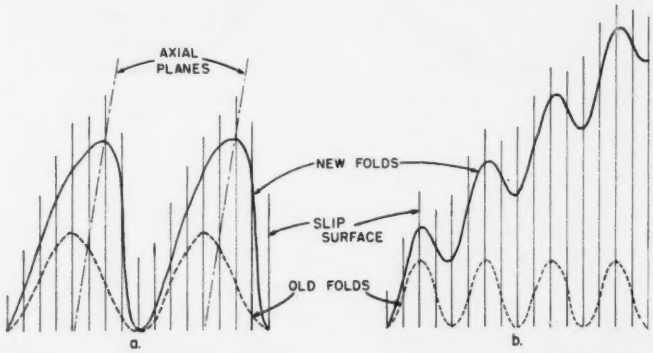

Figure 10. Alternate explanations of the statistical geometry of S_1 and S_2. a. nonaffine slip on S_2 resulting in folds with axial planes inclined to the slip surfaces; b. statistically affine slip on S_2 resulting in folds with the axial planes parallel to the surface of slip, with the west (left hand) limbs lengthened and the east limbs shortened

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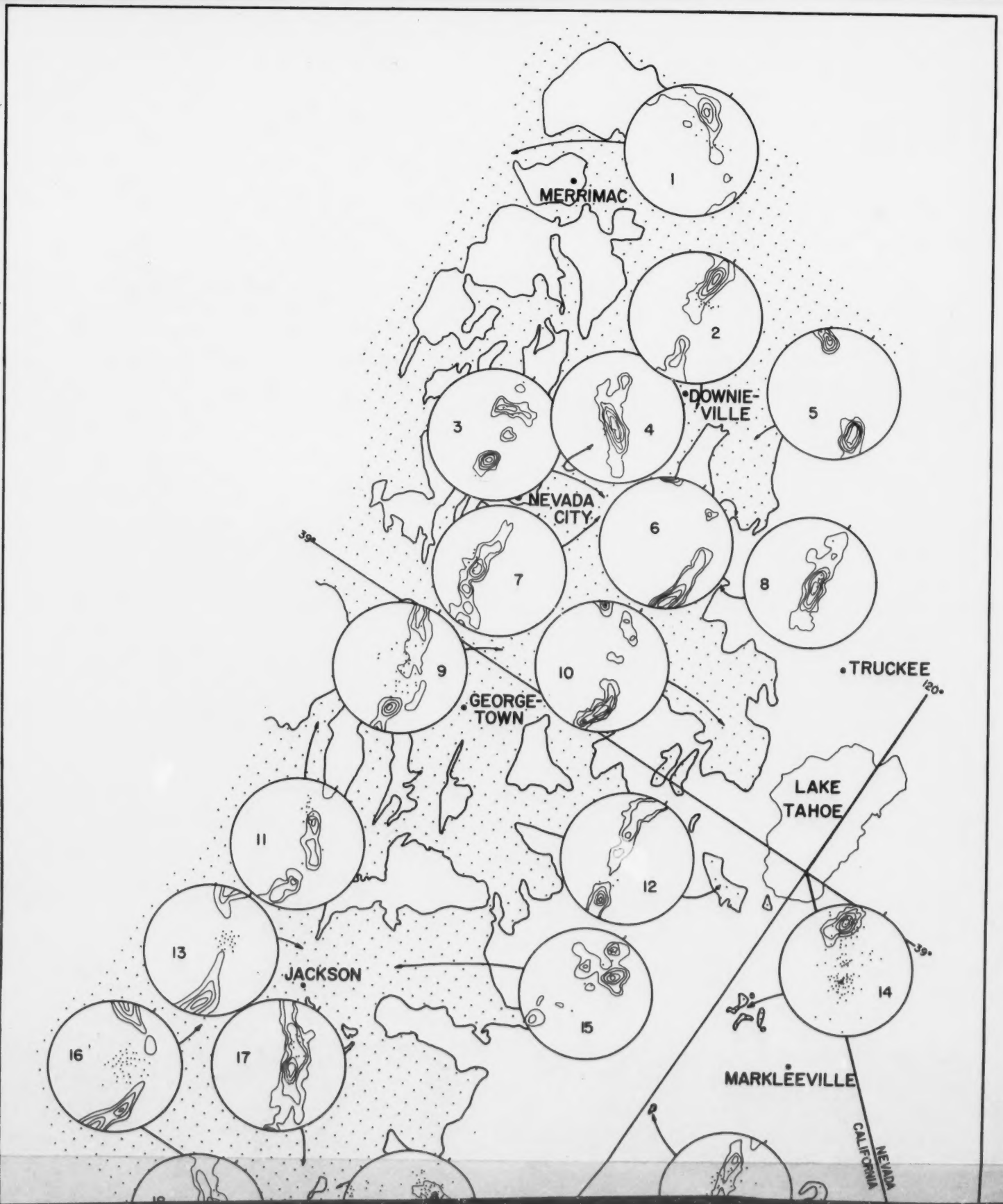
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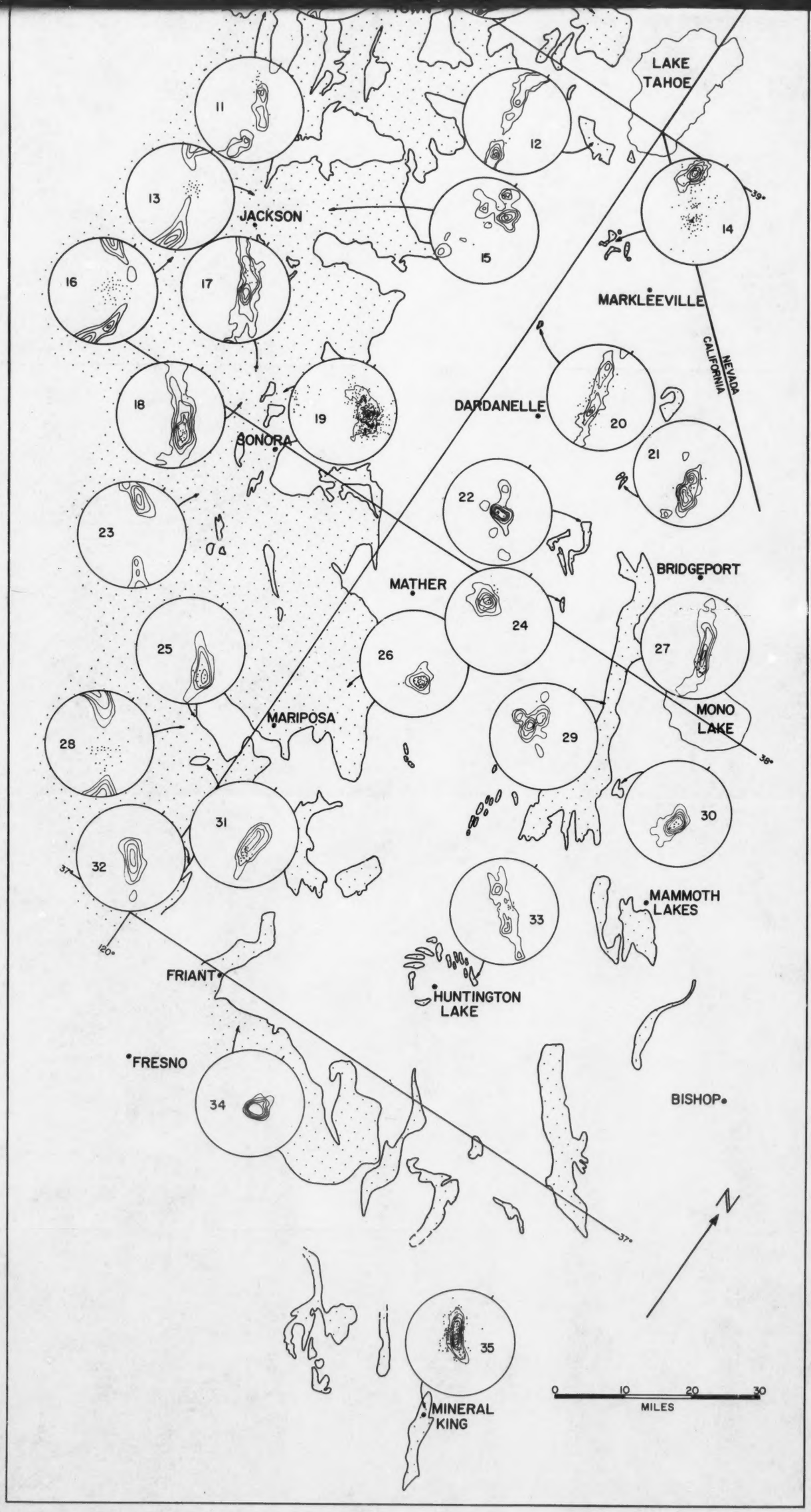

REGIONAL MAP OF THE SIERRA NEVADA SHOWING GEOMETRY OF PENETRATIVE SURFACES AND LINEATIONS IN 35 AREAS

Diagrams are βS contoured at 3–6–9–12–15 per cent per one per cent area (maximum contour less than 15 per cent in some areas). Lineations are shown as points on β diagrams. See appendix for descriptions of areas. Stippled portions are underlain by metamorphic rocks, unstippled areas by granitic rocks. Geology from California State Division of Mines and U. S. Geological Survey maps. Tick marks show true north. Note the preponderance of steeply plunging elements throughout the Sierra Nevada.

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APPENDIX

The following list of sources and types of data are shown diagrammatically in Plate 3. Numbers refer to the numbered β diagrams in Plate 3.

1. Diagram is βS_2; lineations are small fold axes and elongated rock fragments. This area is near a granitic pluton, hence may be affected by forcible intrusion of the granite. The rocks are probably Jurassic. Data include $15S_2$, $7S_1$, 11L (Hietanen, 1951, southwest corner of Pl. 1).
2. Diagram is βS_2; lineations are mineral tracts and axes of kink folds. The rocks are slates of the Calaveras Formation. Data include $12S_2$, 15L (Collected by writer, Highway 49, 1.5 road miles east of Downieville).
3. Diagram is composed of a βS_1 maximum plunging to the north and a βS_2 maximum plunging to the south. The lineations are wrinkles of S_1. The βS_2 is the stronger as in the Alpine County rocks. The rocks are quartzose sandstones assigned to the Blue Canyon Formation by Lindgren (1900). Data include 14S, 6L (Collected by writer, on road from Washington to Relief Hill, near Roscoe Creek).
4. Diagram is βS_2; lineations are wrinkles of S_2. The rock is a massive metasiltstone assigned to the Delhi Formation by Lindgren (1900). Data include $13S_2$, 5L (Collected by writer, southwest of Footes Crossing on North Columbia road).
5. Diagram is βS_1; lineations are S_1/S_2 intersections. Six measurements of S_2 at the same locality define a strong β maximum inclined to the one shown in the diagram. The rocks are bedded cherts assigned to the Calaveras Formation by Turner (1897). Data include $9S_1$, $6S_2$, 2L (Collected by writer, Highway 49, 4.7 road miles west of Sierra City).
6. Diagram is βS_1; no lineations observed. The rock is a bedded calc-silicate hornfels assigned to the Sailor Canyon Formation by Lindgren (1900). Data are $11S_1$ (Collected by writer, 1.0 miles southeast of Bowman Lake).
7. Diagram is βS_2; lineations are elongated pebbles and mineral tracts. The rocks are fissile slates and phyllites with interbedded conglomerate assigned to the Blue Canyon Formation by Lindgren (1900). Data include $12S_2$, 10L (Collected by writer, between Highway 20 and Washington).
8. Diagram is βS (which may be bedding); lineations are mineral tracts and corrugations. The rocks are calc-silicate hornfelses, metavolcanic rocks, and slates assigned to the Sailor Canyon Formation by Lindgren (1900). Data include 13S, 10L (Collected by writer, Highway 40 near east edge of Colfax (30′) quadrangle).
9. Diagram is βS_1; lineations are of many types. Data include $20S_1$, 33L and are from rocks called Calaveras Formation (D. K. Chandra, 1954, Ph.D. thesis, Univ. Calif., Berkeley, Figure 34c, p. 79; p. 148).
10. Diagram is βS_1; no lineations. Data are $11S_1$ from rocks called Milton Formation (Unpublished map by S. G. Clark in files of Univ. Calif., Berkeley).
11. Diagram is βS_1; lineations are mainly mineral tracts. The rocks are probably Jurassic. Data include $18S_1$, 11L (J. F. Crutcher, 1959, M. A. thesis, Univ. Calif., Berkeley, p. 56).
12. Diagram is βS (probably bedding); lineations are small fold axes. The rocks are quartzo-feldspathic hornfelses assigned to the Milton Formation (S. G. Clark, 1930, Ph. D. thesis, Univ. Calif., Berkeley, p. 22–23). Data include 17S, 4L (E. W. Peikert, 1958, M. A. thesis, Univ. Calif., Berkeley, p. 62; written communication, 1959).
13. For this diagram and numbers 16, 23, 25, 26, 28, 31, and 32; the reader is referred to R. J. P. Lyon (1954, Ph.D. thesis, Univ. Calif., Berkeley, p. 86–94) for a discussion of the types of surfaces and lineations.
14. Diagram is βS_1 from subfield II (Fig. 4, this paper) and all lineations from Markleeville quadrangle.
15. Diagram is βS_1; lineations are small folds. The rocks are slates and phyllites, probably Carboniferous. Data include $12S_1$, 12L (Collected by writer, Highway 88, 1.0 road miles east of Jackson).
16. See description 13.
17. Diagram is βS; lineations are schistosity/cleavage intersections and elongated fragments. The rocks are probably in the Calaveras Formation. Data include 19S, 17L (A. K. Baird, written communication, 1959).
18. As in 17. Data include 21S, 28L (A. K. Baird, written communication, 1959).
19. Diagram is βS (said to be possibly relict bedding); lineations are fold axes, wrinkles, grain elongations, and other types. The diagram is generalized from 575 measurements of S, and 486L from rocks (mainly marbles) of Carboniferous age (A. K. Baird, written communication, 1959).
20. Diagram is βS_2. The rocks are of unknown age. Data include $20S_2$, 35L (Gilbert, 1959, M.A. thesis, Univ. Calif., Berkeley, Figure 16B; p. 68).
21. Diagram is βS_2; lineations are elongated pebbles and mineral tracts. The β axis defined by six measurements of S_1 is inclined 12° to the βS_2 axis. Data include $12S_2$, 11L (Collected by writer, near Leavitt Meadows).
22. Diagram is βS_2; lineations are mainly corrugations. The rocks are Mesozoic metavolcanic rocks. Data include $12S_2$, 7L (Clyde Wahrhaftig, personal communication, 1959, NE corner Tower Peak quadrangle).
23. See description 13.
24. Diagram is βS_1; lineations are mainly corrugations. The rocks are hornfelses of unknown age. Data include $20S_1$, 20L (R. L. Rose, 1957, Ph. D. thesis, Univ. Calif., Berkeley, p. 219).
25. See description 13.
26. See description 13.
27. Diagram is βS_1; lineations are mainly corruga-

tions. The rocks are hornfelses of unknown age. Data include $18S_1$, 20L (Stanford University Geology Summer Camp, map of east North Peak area, made available by G. A. Davis).
28. See description 13.
29. Diagram is βS_1; lineations are mainly corrugations. The rocks are hornfelses of unknown age. Data include $14S_1$, 6L (R. W. Kistler, personal communication, 1959, northeast of Gem Pass, Mono Craters quadrangle).
30. As in 29. Data include $20S_1$, 8L (R. W. Kistler, personal communication, 1959, east of Silver Lake, Mono Craters quadrangle).
31. See description 13.
32. See description 13?
33. Diagram is βS; lineations are corrugations and small fold axes. The rocks are marbles of possible Carboniferous age. Data, include 22S, 14L (Collected by writer, roof pendant near Twin Lakes, Fresno County).
34. Diagram is βS. The rocks are metavolcanic rocks of late Paleozoic or early Mesozoic age. Data include 20S (Macdonald, 1941, map).
35. Diagram is βS, lineations are of several types. The rocks are hornfelses, schists, and marbles of Mesozoic age. Data include 215L and a synoptic β diagram (M. L. Christensen, personal communication, 1959).

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K-Ar Age Studies of Mississippi and Other River Sediments

Abstract: The ratio of radiogenic argon 40 to potassium supplies information on the source or genesis of components of recent sediments. This is the initial investigation in a larger study on the history of illitic materials using dating techniques. In samples from the deltaic deposits of the Mississippi River, age values for the dominantly silt-sized material of the delta averaged 280 m.y., with little horizontal or vertical variation. The clay-size fractions of the same samples averaged only 166 m.y. This age is lower than one would expect of detritus from Paleozoic shales in the central and eastern part of the source area but may be due to mixtures of younger mixed-layer illite-montmorillonite released from the Cretaceous and Tertiary sections to the west.

The Mississippi River sediments in the delta are commonly gray. A reddish-brown layer found at depth in drill holes, believed to have been deposited by the Red River, showed ages of 690 m.y. for the dominantly silt-sized whole samples and 650 m.y. for the clay fractions. This supports a Red River source for this layer and suggests a common source for the two fractions. One would expect the rapidly deposited deltaic sediments from both the Mississippi and Red rivers to lack any appreciable authigenic K phases, if such exist, because these require slow accumulation well offshore.

Measured ages of the clay fractions in bottom sediments from the Rappahannock River in Virginia are similar to those from the crystalline rocks of the source regions, indicating again that the K phases are ancient detritus.

Clay soils forming over ancient shales can contain illite of the same age as the underlying shales, thus indicating a source for detrital, ancient $1M_d$ illite.

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INTRODUCTION

The relationship of potassium to argon in K-bearing minerals has been applied almost exclusively to investigations of methods of age dating. It can also be used for other purposes, and the authors are studying the genesis, history, and patterns of migration of potassium-bearing clay minerals by this means.

The subject of clay-mineral genesis has received much attention lately, and as this paper introduces another area of research with its

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own large bibliography, the writers believe it advisable to refer the reader to comprehensive papers in both areas. A comprehensive paper by Weaver (1959) on the clay petrology of sediments supplies many of the data and ideas utilized in the present paper. Papers by Tilton and Davis (1959) and Aldrich and Wetherill (1958) cover the bibliography and scope of the investigations involving K-Ar dating. Hurley, Cormier, Hower, Fairbairn, and Pinson (1960), in a precursor to the present investigation, report on the measurement of the age of glauconites.

The principal potassium-bearing detrital phases from crystalline source rocks will be micas and feldspar, and the dioctahedral mica will be the 2M type. Current hypotheses state that, in the clay-size fractions in shales, the $2M_1$ muscovite will be detrital, but the $1M_d$ phases may be either detrital or syngenetic. Also, epigenesis may occur upon burial, involving the further development of $1M_d$ illite or interlayered illite-montmorillonite. Source rocks for ancient detrital $1M_d$ illites would presumably be ancient unmetamorphosed sediments. Both $1M_d$ and $2M_1$ polymorphs may be partly stripped of potassium during weathering, so that ancient basic structures may be available for potassium reconstitution in marine environments. In this case the material in the sediment would show a reduced age representing the original K-Ar age ratio in the unstripped part diluted by the added potassium. Syngenesis of illite or interlayered illite-montmorillonite from beidellitic or other material may occur in the marine environment, but the dominant process of formation of zero-age illite is believed to be the cation reconstitution of original basic structures, or true crystal growth in the sedimentary environment.

The writers are investigating K-Ar age ratios in separated $2M_1$ and $1M_d$ components in illitic shales of various ages, in soils, and in deep ocean sediments. Each investigation is planned to aim at a single question, and will be reported separately. This part of the study is on the detrital sediments at the mouth of a river, with little or no chance of authigenesis or modification owing to active deposition.

ACKNOWLEDGMENTS

The authors are particularly indebted to H. N. Fisk and the Humble Oil and Refining Company for supplying the samples, drill logs, and X-ray analytical data for the materials from the Mississippi delta. B. W. Nelson of the Virginia Polytechnic Institute very kindly donated the samples of bottom sediments from the Rappahannock River. J. Hower, C. E. Weaver, H. N. Fisk and B. W. Nelson made many helpful suggestions, and H. W. Krueger assisted with some of the analyses. The authors wish to thank all of these collaborators, and also to express appreciation for support by the Division of Research of the U. S. Atomic Energy Commission.

MISSISSIPPI DELTAIC DEPOSITS

In this study the writers were pleased to have the cooperation of Harold N. Fisk who has studied in detail the materials of the Mississippi Delta (Fisk and McClelland, 1959; Fisk and McFarlan, 1955). His reports and that of Taggart and Kaiser (1960) are pertinent because they were based in part on the same surface and drill-hole samples used in this investigation, and thus provide detailed information on these materials.

The present deltaic sediments of the Mississippi River were deposited following the peak of the last glaciation. Radiocarbon dating (Brannon *et al.*, 1957) has established that, approximately 40,000 years ago, sea level was about 450 feet lower than at present. The Mississippi River at that time flowed through a trench cut into a weathered, oxidized Pleistocene surface, establishing new deltas which now appear as shoals on the continental shelf. The rise in sea level caused a filling in of the trench, accompanied by compaction and downwarping of the sediments and continued migration of the river. Fisk and associates have traced the old river courses and have outlined the complex of deltas in the present deltaic plain. More than 8000 cubic miles of sediments has been deposited by the river in a 44,000-square-mile area, causing a troughlike downwarping in the continental margin of 350 feet near the present shore line and 500 feet on the shelf.

Most of these sediments have accumulated at the delta front and on the deeper gulf floor surrounding it, *i.e.*, the prodelta marine environment. Most of the shallow (less than 60 feet of water) delta-front facies are massive silty clays. The deeper-water deposits are finely laminated. The marsh deposits of the delta plain consist mostly of massive silty clays containing roots and local peat beds, whereas more sandy facies occur in lakes and bays, beaches, and offshore bars. Although the Mississippi River has been the principal source, there are indications that the Red River has deposited

sediments locally. The Mississippi River sediments are characteristically grayish, whereas clays from the Red River are commonly reddish-brown.

Taggart and Kaiser (1960) studied the clay mineralogy. They conclude that, on the basis of abundance, variation in montmorillonite content was the most outstanding and definite difference. A preponderant number, 47 of the 60 deltaic samples, had montmorillonite as the most abundant clay mineral, followed by illite, kaolinite, and chlorite. On the other hand, the Red River sediments contained little or no montmorillonite. Their results support the conclusions that montmorillonite does not alter to form illite or chlorite in the marine environment, and that the clay-mineral composition of source sediments determines the clay-mineral content of marine sedimentary rocks.

SAMPLES ANALYZED

The locations of the bottom samples are shown on Figure 1. Eight samples are of offshore marine delta front and prodelta clays which underlie the Gulf floor at depth ranging from 0 to 447 feet below the solid bottom surface. Five of these samples are from a single boring located near the mouth of South Pass, and two others are from a second boring off Jefferson Parish. Lithologic and paleontologic characteristics indicate the stratigraphic positions (Fig. 2). The remaining five samples are surface or near-surface materials from the deltaic plain. These samples represent materials from fresh-water levee and marsh, brackish-water marsh, and salt-water marsh (Table 1).

Samples for analysis were dried, dispersed by

Figure 1. Location of samples from Mississippi delta

ultrasonics in water without dispersing agent, and centrifuged. Owing to the small amount of original sample, the clay-sized fraction was taken to be −4 microns. Sample S-3998 silt was further separated into magnetic and nonmagnetic fractions in a Frantz separator.

is third in abundance, and in S-4002 and S-4003 the order of illite and kaolinite is reversed. All samples contain quartz and feldspar. Although the original whole-sediment samples correspond with those measured for potassium and argon, the clay-size fractions were prepared

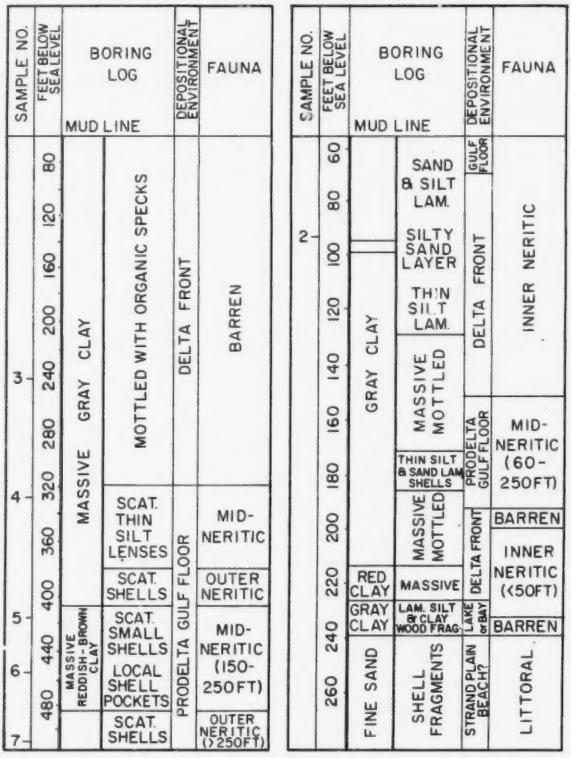

Figure 2. Logs of bore holes in Mississippi delta, showing sample locations

X-RAY ANALYSES

Information on the relative abundance of clay minerals in the samples has been kindly provided by the Humble Oil and Refining Company. The analyses were made under the direction of M. S. Taggart and A. D. Kaiser, Jr. (1960, p. 524–526). The results indicate that except in samples S-4001, S-4002, and S-4003, the order of decreasing abundance is montmorillonite, illite, kaolinite, chlorite. In S-4001 kaolinite is most abundant and montmorillonite

independently. Most of the samples were dominantly of silt-size material, with relatively little clay. The coarser material was mostly quartz, with some feldspar.

ANALYTICAL RESULTS

Analytical techniques were essentially the same as those described by Hurley *et al.*, (1960) except that a small amount of NaOH flux was used, and the gas analyzed exclusively in a Reynolds-type mass spectrometer (Reynolds, 1954).

In analyses with air corrections less than 30 per cent the estimated precision (reproducibility) was 3 per cent standard deviation. Air corrections above this figure cause the precision error to rise, for although the error in 36/38 measurement is fixed, it enters as the difference in subtracting the air Ar^{40}. The writers believe systematic error of the mean to be less than 1 per cent, based on the measurement of the argon content of air, on spectroscopically pure argon, and on interlaboratory agreement in the analysis of a standard sample of biotite (B-3203).

Decay constants for potassium-40 have been adopted as follows: $\lambda_e = 0.585 \times 10^{-10}$ yr.$^{-1}$; $\lambda = 5.30 \times 10^{-10}$ yr.$^{-1}$ (Aldrich and Wetherill, 1958).

K-Ar age ratios on the samples of whole sediment from the delta-plain deposits and from the prodelta deposits are listed in Tables 2 and 3. The principal K-bearing components are detrital, and at least Paleozoic in age.

The coarser material of the clay-size fractions consists of terrigenous minerals, plant fibers, and, in some, shells, foraminifera, and detrital glauconite. Shepard and Lankford (1959) have

Table 1. Description of Samples

Sample number on map	M.I.T. sample number	Description
1	S-3997	Gulf floor marine clay, 40.5 feet below sea level and 1.5 feet below Gulf bottom; core from Humble Oil and Refining Co. well F-1, State Lease 799, off Caminada Pass, 29°08.6′N. latitude, 90°0.10′W. longitude, off Lafourche Parish, Louisiana
2	S-3998	Marine Clay from delta front deposits, 97 feet below sea level and 42 feet below Gulf bottom; core from Humble Oil and Refining Co. well No. 1-J, Louisiana State Lease 804, Grand Isle Block 23, 29°06.1′N. latitude, 89°59.5′W. longitude, off Jefferson Parish, Louisiana
3	2-3999	Marine clay from delta front deposits, 246 feet below sea level and 187 feet below Gulf bottom; core from Humble Oil and Refining Co. well No. 1, Louisiana State Lease 2729, South Pass block 20, 29°00.0′N. latitude, 89°07.1′W. longitude, off Plaquemines Parish, Louisiana
4	2-4000	Marine clay from prodelta deposits, 326 feet below sea level and 267 feet below Gulf bottom; core from same well and location as sample No. 3
5	S-4001	Marine clay from prodelta deposits, 416 feet below sea level and 357 feet below Gulf bottom; core from same well and location as sample No. 3
6	S-4002	Marine clay from prodelta deposits, 456 feet below sea level and 397 feet below Gulf bottom; core from same well and location as sample No. 3
7	S-4003	Marine clay from prodelta deposits, 506 feet below sea level and 447 feet below Gulf bottom; core from same location as sample No. 3
8	S-4004	Silty organic clay from beneath peaty clay of salt-water marsh deposit; sample is from cuttings 7.5 to 9.0 feet below sea level and 6.0 to 7.5 feet below ground surface, Peat Boring 102, 4,200 feet south and 2,000 feet east of N.W. corner of sec. 15, T. 20 S., R. 28 E., about 5 miles west of Mississippi River near Empire, Plaquemines Parish, Louisiana
9	S-4005	Clay from bay deposits in salt-water marsh 5.5 to 6.0 feet below sea level and 0.5 to 1.0 feet below bottom of Hackberry Bay; core from Marsh Hydrographic Station 16, 29°25′51″N. latitude, 90°01.39′W. longitude, Lafourche Parish, Louisiana
10	S-4006	Organic clay beneath surface vegetal muck from brackish-water marsh 2.5 to 3.0 feet below sea level and 2.5 to 3.0 feet below ground surface; core from Peat Boring 99, 800 feet north and 2,100 feet east of SW. corner sec. 33, T. 15 S., R. 15 E., Plaquemines Parish, Louisiana
11	S-4007	Clay from Bayou des Alemands located in fresh-water marsh, 7.25 to 8.0 feet below sea level and 0.25 to 1.0 feet below bayou bottom; core from Marsh Hydrographic Station 1A, 29°46′05″N. latitude, 90°23′40″W. longitude, Lafourche Parish, Louisiana
12	S-4008	Fresh-water silty clay from Mississippi River levee 13.4 to 13.7 feet above sea level and 1.3 to 1.6 feet below levee surface; core from Peat Boring 36 at Laplace, Louisiana, 3,000 feet south and 8,000 feet west of NE. corner sec. 27, T. 11 S., R. 7 E., St. John the Baptist Parish, Louisiana
2	S-4013	Reddish-brown clay from 217.5 feet below sea level and 162.5 feet below Gulf bottom in Humble Oil and Refining Co. well No. 1-J, Grand Isle Block 23, 29°06.1′N. latitude, 89°59.5′W. longitude

TABLE 2. ANALYSES OF WHOLE SEDIMENT SAMPLES

Sample number	Depth below solid surface	Location, parish	%K	Radiogenic Ar^{40}/K^{40}	Air correction	Age value m.y.
Deltaic Plain						
Salt- and Brackish-Water Marsh Deposits						
S-4004	7	Plaquemines	2.13	.0184	24	290
S-4005	1	Lafourche	2.22	.0150	16	240
S-4005	(Acid treated)		2.11	.0155	51	247
S-4006	3	Plaquemines	2.46	.0133	33	214
Fresh-Water Marsh Deposits						
S-4007	1	Lafourche	2.16	.0144	41	230
S-4008	1.6	St. John the Baptist	2.05	.0230	15	357
Prodelta Deposits						
Gulf-Floor Marine Clay						
S-3997	1.5	Lafourche	2.40	.0195	(30)%	(306)
Marine Clay from Delta-Front Deposits						
S-3998	42	Jefferson	1.78	.0213	15	333
S-3999	187	Plaquemines	2.13	.0183	15	289

analyzed the components in several shallow borings in the delta and prodelta deposits. The terrigenous-mineral component is dominated by quartz, but some orthoclase, mica, and glauconite will undoubtedly affect the age ratios. Table 4 is a comparison of the magnetic versus nonmagnetic sand-size material, as separated by Frantz magnetic separator. The mafic component shows a high potassium content, explained by the presence of biotite and rounded grains of glauconite. The nonmagnetic fraction has less potassium because of its quartz content, but the principal K-bearing phases are probably accounted for by microscopically determined orthoclase and mica.

Analytical results are given on a group of core samples from a deep drill hole in the prodelta sediments (Table 3). The gray sediments, presumably of Mississippi River origin, show about the same age as do those in Table 1, but the reddish-brown sediments, believed to be of Red River origin (Taggart and Kaiser, 1960), show a much greater age, indicating that the detrital K-bearing components were from a Precambrian crystalline source. The writers suggest the Rocky Mountain Front

TABLE 3. ANALYSES OF WHOLE SEDIMENT SAMPLES FROM DRILL HOLES

Sample number	Origin and environment	Depth below solid surface	Color	%K	Radiogenic Ar^{40}/K^{40}	Air correction	Age value m.y.
Humble Oil and Refining Co. Well No. 1, South Pass Lat. 20°—'N, Long. 89°07.1'W							
S-3999	Miss. R., Delta Front	187	Gray	2.13	.0183	15%	289
S-4000	Miss. R., Gulf Floor	267	Gray	2.25	.0254	23	390
S-4001	Red River (?)	357	Red-brown	2.54	.0378	8	540
S-4002	Red River (?)	398	Red-brown	2.52	.0578	6	795
S-4002	repeat	398	Red-brown	2.52	.0580	6	795
S-4003	Miss. R., Gulf Floor	447	Gray	2.35	.0194	24	303
S-4003	repeat	447	Gray	2.35	.0177	31	280
Humble Oil and Refining Co. Well No. 1-J, Grand Isle Lat. 29°06'N, Long. 89°59.5'W							
S-3998	Miss. R., Delta Front	42	Gray	1.78	.0213	15	333
S-4013	Red River (?)	162	Red	2.65	.0592	7.2	810
S-4013	(+4 size only)*	162	Red	2.51	.0725	7.9	950

* The +4 size fraction represented only 12.5 per cent of the sample by weight. The −4 fraction analysis is given in Table 3.

TABLE 4. COMPARISONS OF VARIOUS SIZE FRACTIONS WITH WHOLE SAMPLES

Sample number	Whole sample %K	Whole sample Age, m.y.	Weight %	Size	%K	Ar^{40}/K^{40}	Air corr. %	Age value m.y.
Mississippi River Gray Sediment								
S-3998	1.78	333	7	-2μ	2.18	.0062	60	102
			33	$+2\ \mu$ -400 mesh	2.21	.0161	6	256
			10	+ 400 mesh magnetic	2.45	.0226	30	336
			50	+400 mesh nonmagnetic	1.47	.0265	17	405
S-3997	2.40	306		$-4\ \mu$	2.14	.0107	57	175
S-4003	2.35	290 (Av.)		$-4\ \mu$	2.61	.0128	54	206
S-4004	2.04	302		$-4\ \mu$	2.35	.0102	42	166
S-4005	2.22	240		$-2\ \mu$	2.49	.0098	41	160
S-4005 Acid Treated	2.11	247		$-4\ \mu$	2.51	.0088	55	145
S-4006	2.46	214		$-4\ \mu$	2.36	.0109	43	178
S-4007	2.16	230		$-4\ \mu$	2.29	.0097	44	160
S-4008	2.05	357		$-4\ \mu$	2.26	.0128	44	206
Red River(?) Reddish-Brown Sediment								
S-4001	2.54	540		$-4\ \mu$	3.07	.0322	17	483
S-4013	2.53	840		$-4\ \mu$	2.51	.061	7	830

Range, as 800 m.y. would represent an average value for detrital feldspars from the area, although it would be a little low for micas. Tests of reddish-brown samples from two drill holes, far apart in the delta, gave almost identical results.

Clay-size fractions separated from the above samples are listed in Table 4. Again the most obvious observation is the difference between the gray and reddish-brown samples, the Red River material being more than twice the age of that from the Mississippi River.

A single test of an acid-treated sample (S-4005A) in both whole sediment sample and clay fraction shows no obvious effect of such treatment either in the potassium content or age value (Table 4).

CLAY SOILS FORMING ON ANCIENT ILLITIC SEDIMENTS

The natural consequence of the previous analyses was to determine whether ancient illite was released from underlying shales during weathering, without loss of radiogenic argon. The writers selected a sample of illitic soil from an area in northern New York state in which Devonian illitic shale was decomposing into a soil containing illite. The results are shown in Table 5.

It is clear that the illite in the soil is the same

TABLE 5. AGE OF ILLITE IN CLAY SOILS DERIVED FROM ANCIENT ILLITIC SOURCE ROCKS

Sample number	Sample locality and description	%K	*Ar^{40}/K^{40}	% Air correction	Age m.y.
R 3898	-1μ illite-chlorite fraction, from Devonian shale, Truxton, N.Y.	4.81	.0207	24	325 ± 30*
R 3901	-1μ illite-chlorite from soil near Aurora, N.Y. Overlying Devonian shale	2.33	.0215	28	336
R 3899	Vicksburg, Miss. loess whole soil, 30–40% illite-montmorillonoid clay	1.71	.0315	3	474

* The estimated analytical error for these clay materials is much higher than for clean mica of larger grain size. The cause of the poorer reproducibility in analysis is not known, but may be connected with weakly held radiogenic argon. The error rises sharply with the air correction.

age as the illite in the underlying rock. The K-Ar age in both is slightly less than the age of the sedimentation. It is probable, however, that some potassium and argon are removed equally during weathering. This suggests that much wind-borne dust may be composed of ancient micas, derived from the unlocking of the clay-size particles in shaly sediments. The show mineral ages corresponding to the Taconic, Acadian, and Appalachian orogenic events, which roughly fall in the age range of the up-river samples. Sediments from the estuary may originate in part from ocean currents, in part from other sources—some, at least, Precambrian. Therefore, the detrital K-bearing components in the mouth of the

TABLE 6. CORE SAMPLES FROM THE BOTTOM OF THE RAPPAHANNOCK RIVER, VIRGINIA
(Prepared and donated by Bruce Nelson)

Sample number	Distance from river mouth	Depth in core, inches	Size fraction	%K	Radiogenic Ar^{40}/K^{40}	Air correction	Age value m.y.
S-3863	68 miles	2–4	-4μ	1.49	.0214	55	334
S-3865	68	11–13	-4μ	1.47	.0184	32	290
S-3866	64	11–15	-4μ	1.46	.0304	44	460
S-3867	2	2–9	$4-15\mu$	1.94	.0480	35	680
S-3868	2	2–9	-4μ	2.32	.0333	12	497

writers made a test on a sample of Mississippi loess (Table 5). Details are as follows:

R-3899, Vicksburg loess. Illite-montmorillonoid clay from Vicksburg, Miss., from height of land above the river cut. Supplied by Waterways Experimental Station, Vicksburg. Sample of whole soil containing 30–40% clay mineral, 15% carbonate, and 0.5% organic matter. (Sample provided by R. T. Martin, Mass. Inst. Technology).

SAMPLES OF BOTTOM MUDS FROM THE RAPPAHANNOCK RIVER IN VIRGINIA

No samples were obtained of the suspended materials in the Mississippi River away from its mouth. The writers were pleased to receive some bottom-sediment samples from the Rappahannock River, Virginia, from B. W. Nelson of Virginia Polytechnic Institute. These samples were part of an extensive study of the clay mineralogy of the bottom sediments in this river (Nelson, 1960), in which progressive changes were observed between the fresh-water and saline-water portions of the Rappahannock system.

Analytical results of three fresh-water samples and two marine samples are given in Table 6. Here also the K-bearing phases appear to be detrital. The region being drained by the river is underlain by crystalline rocks that Rappahannock River probably have been modified by long-shore currents and atmospheric dust. Certainly the question of authigenesis of the illitic material in the estuarine deposits must be examined in the light of these results.

DISCUSSION OF RESULTS AND CONCLUSIONS

The coarser-grained material bearing potassium is dominantly in the age range characterized by most of the minerals found in the crystalline rocks of the Appalachian Mountain system, drained to the west by the tributaries of the Ohio River. This age (about 400 m.y.) is lowered in the whole sample by the younger clay-sized materials. The age for the magnetic fraction, similar to the nonmagnetic, is dominated by biotite from the same source, modified slightly by glauconite which is dominantly of mixed Paleozoic age, and could have an average age close to this value. All other mineral and nonmineral components act as diluents to the K-bearing phases and do not affect the age values. The delta and prodelta whole samples from many locations and environments all show a rather similar age.

Sediments from the Red River, although interbedded with those of the Mississippi at depth in the delta, show a very different age, reflecting a Precambrian source. The clay-size fraction shows about the same age as the coarse fraction and is probably the finer sizes of material from the same source. As it is Pre-

cambrian in age, the muscovite is presumably of the $2M_1$ type since any subsequent modification would have lowered the observed age.

The sketchy data on the Rappahannock bottom sediments indicate that at least most of the K-bearing clay-size material is detrital, including the increasingly illitic clays obtained from the marine environment in the estuary.

The age of the clay fractions in the Mississippi sediments is much less than that of the coarser fractions. Johns and Grim (1958) have concluded that the Missouri River contributes mostly montmorillonite of bentonitic origin, whereas the Ohio River contributes much more illite. The illitic detrital component in the delta and prodelta sediments therefore probably comes principally from an area underlain by Upper Paleozoic shales. The clay fractions from such shales and also from soils overlying them show K-Ar ages in the range of 250–350 m.y. for material of Carboniferous through Devonian age. The observed ages in Table 4, mostly in the range 150–200 m.y., are lower by a factor of nearly two. The writers conclude that the detrital illitic material in these samples is diluted by material that has a lower argon age. Furthermore, since the detrital illitic component probably contains much more potassium than the more abundant diluting material, it dominates the age values disproportionately.

C. E. Weaver (Personal communication) suggests that this lower-age potassium-bearing diluting material may be the montmorillonite or mixed-layer illite-montmorillonite, which has been largely contributed to the river by the drainage system to the west. He states that this montmorillonite comes mainly from the Cretaceous and Tertiary areas and is therefore of young age, and that it commonly carries 2–4 per cent K_2O. If this is true, the young age of the aggregate would be the result of mixing two detrital materials, each reflecting the age of its source region. The writers agree with Weaver and simply add that if there is also $1M_d$ illite of low age from the Paleozoic shales, this would contribute to the reduction in age of the mixed aggregate. It is most unlikely that any significant development of new K-bearing phases would occur in the delta deposits, as they are made up of very rapid accumulations of mostly silt-sized material. One would expect significant amounts of authigenic clay minerals only in quiet areas of slow deposition far from shore.

The clay fractions from the Mississippi sediments all show large air corrections. These materials apparently have a strong affinity for atmospheric gases. Atmospheric argon-40 continues to be released strongly when the sample is heated to the point at which radiogenic argon is given off. Also the material from fresh- and brackish-water environments shows the same lower ages and high air-contamination corrections. The writers have analysed the $1M_d$ fraction separated by S. W. Bailey from a Pennsylvanian shale and found the same characteristics. On the other hand, the Red River clay material (S-4001 and S-4013 in Table 3) has much less chemisorbed atmospheric gas and shows a higher age. This may well be because it is dominantly $2M_1$ mica. The clay fractions from river and ocean-bottom sediments, soils, shales, and wind-borne dust all show less tendency for strong adsorption or occlusion of atmospheric gases.

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Chemical and Mineralogical Composition of Fossil Molluscan Shells from the Fox Hills Formation, South Dakota

Abstract: Chemical composition and mineralogy of a large number of samples of molluscan shell material from limy concretions in the Fox Hills Formation (Cretaceous) of South Dakota were determined. The results show an extremely wide range of concentrations for the minor elements Mg, Fe, Mn, Sr, and Ba and of calcite-aragonite ratios.

Detailed studies of an ammonite, *Sphenodiscus*, a characteristic faunal element of this formation, showed that the range of minor-element concentrations could best be interpreted as the consequence of at least two separate alteration processes, involving contamination and recrystallization, probably during diagenesis. Neither visual nor mineralogic criteria can be used to predict the degree of alteration of the original shell composition. Since the alteration processes may have both added and subtracted material, one must be cautious in accepting any fossil-shell-composition data as representative of the originally deposited test.

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INTRODUCTION

Schemes to decipher ancient environments from the mineralogical or chemical composition of fossils or associated sedimentary rocks have aroused considerable interest (Urey *et al.*, 1951; Kulp, Turekian, and Boyd, 1952; Chave, 1954a; 1954b; Lowenstam, 1954, 1961; Turekian, 1955; Odum, 1957). The conditions for success are stringent, requiring both some parameter that is unequivocally sensitive to environment and a guarantee that no significant alterations of this information have occurred since sediment and shell were removed from the contemporary environment.

The sight of light-colored iridescent shell material preserved in calcareous concretions in the Fox Hills Formation (Cretaceous) of South Dakota suggested the possibility of measuring original chemical and mineralogical properties

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of these shells. The Fox Hills Formation was deposited during the retreat of the last major epeiric sea on the North American continent; it records a continually changing environment from marine to near-continental.

To understand the composition of ancient molluscan tests (Turekian and Armstrong, 1960).

The most general change that takes place during fossilization and diagenesis is apparently the loss of strontium from the carbonates (Kulp, Turekian, and Boyd, 1952; Siegel,

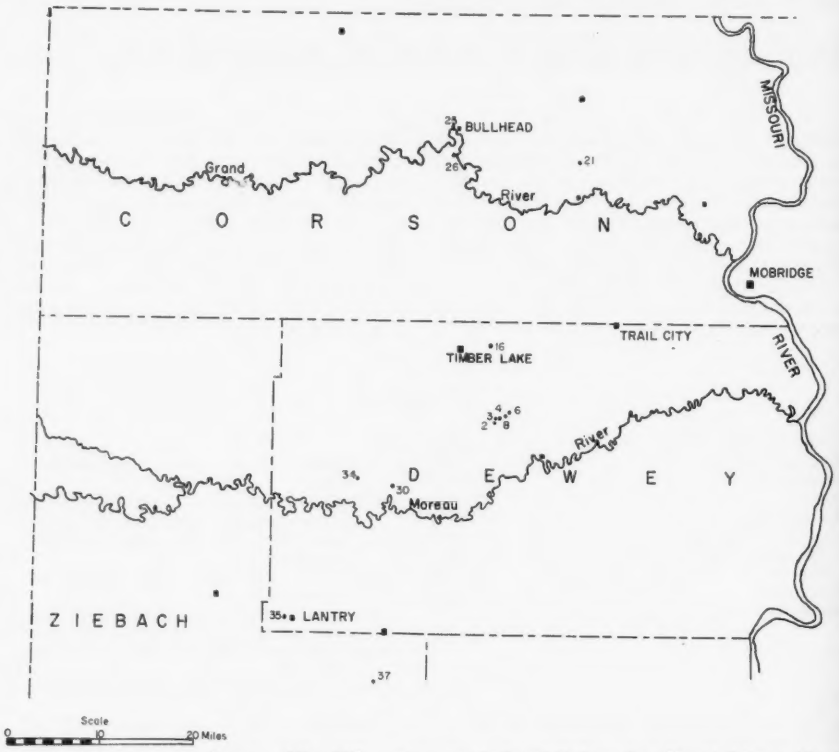

Figure 1. Location map of region of Fox Hills outcrop area in South Dakota. Locality 44; Linton, North Dakota; and Pierre shale locality near Wasta, South Dakota, not on map

shells it is important to have information on the closest contemporary representatives. The Fox Hills fauna is mainly molluscan with a strong ammonite component. As this group is extinct, we must make comparisons with other living mollusks from a variety of environments and hope that these may be adequate contemporary chemical and mineralogical representatives of the ancient fauna. We have made such a study on the magnesium, strontium, and barium concentrations and the calcite-aragonite ratios in a large number of Recent marine

1960). The behavior of barium and magnesium (Krinsley, 1960) is less predictable, as it is strongly controlled by sedimentary environment.

If it can be demonstrated that mineralogically unaltered shells retaining their original microscopic structure can undergo trace-chemical changes during diagenesis, then any study depending upon retention of original composition must demonstrate the absence of alteration by other means. This condition encompasses not only studies of ancient environ-

ments but also radioactive dating techniques based on the uranium-decay series, as suggested by W. M. Sackett (1958, Ph.D. thesis, Washington Univ.) and W. S. Broecker (Personal communication, 1960). The possibility of the loss or gain of any of the daughter products cannot be ignored, and suitable tests must be devised to ensure that the system has been closed since burial or that all diagenetic changes involving these elements took place very rapidly after burial after which the system remained essentially closed.

METHODS

Figure 1 is a map showing the localities of the Fox Hills Formation sampled by Waagé and studied chemically and mineralogically by us. Waagé (1960; 1961) has discussed the faunal distribution in the Fox Hills Formation of South Dakota, and Table 1, is based on his work.

TABLE 1. GENERALIZED SECTION SHOWING CONCRETIONARY ZONES
(Waagé, personal communication, 1960)

Formation		Fossils
	Hell Creek Formation Colgate Member	Concretion zones
Fox Hills Formation	Bullhead Member	
	Timber Lake Member	← Oyster ← *Pteria linguiformis*
	Trail City Member	← *Protocardia* ← *Limopsis—Gervillia* ← Lower *Discoscaphites nicollets*
Pierre Shale	Elk Butte Member	

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The fossil shells used in this study were mainly removed from limy concretionary material. The flakes of shell were carefully separated both from the matrix of the concretion and obvious secondary calcite crystals. Samples of the concretion matrix were also collected and analyzed; results on these will be discussed elsewhere.

The emission spectrographic technique for the determination of magnesium, strontium, and barium and the X-ray diffraction techniques for the determination of the calcite-aragonite ratio have already been described (Turekian and Armstrong, 1960).

Manganese and iron were also determined from the same plates used for the magnesium determinations. The Mn 2801.06 and Fe 3008.14 lines were compared to the Ca 3009.20 line. Standards were made by mixing known

TABLE 2. CHEMICAL COMPOSITION AND MINERALOGY OF SHELLS OF REPRESENTATIVE FAUNA FROM THE UPPERMOST PIERRE SHALE AND LOWER FOX HILLS FORMATION

Species	Class*	Per cent calcite	Mg	Sr	Ba
PIERRE SHALE ("SAGE CREEK") ZONE					
Locality: near Wasta, South Dakota					
Eutrephoceras	C	1.	375	3350	95
Placenticeras	C	1.	260	3400	105
Acanthoscaphites	C	2.	905	2850	720
Inoceramus	P	0.5	105	3250	190
Concretion	..	64	20,800	..	..
FOX HILLS FORMATION (TRAIL CITY MEMBER) Lower *Discoscaphites nicolleti* layer					
Locality 2					
Discoscaphites nicolleti	C	1	270	> 5000	740
Goniomya americana	P	1.5	325	1700	155
Unidentified snail	G	18	940	2800	560
Inoceramus	P	3	500	3150	765
Concretion	..	22	16,500	1750	835
Locality 44					
D. nicolleti	C	10	1100	5000	405
Unidentified snail (chalky)	G	18	660	2450	280
Thracia subtortuosa (?)	G	86	2450	1175	150
Tellina	P	10	560	2150	165
Protocardia subquadrata	P	30	1525	..	..
Concretion	..	22	18,000	..	..
Locality 21					
Gervilla	P	0.5	145	..	..
Nucula (?)	G	8	1600	..	..
Concretion	..	22	12,000	..	..
D. nicolleti	C	1	660	..	..
Inoceramus	P	8	1850	..	..
Snail (unidentified)	G	3	660	..	..
***Limopsis—Gervillia* layer**					
Locality 3					
L. striata Kuncatata	P	3	560	2500	800
Sphenodiscus	C	25	132	2450	> 2000
Nautiloid	C	37	730	2300	170
D. Cheyennensis	C	0	145	3100	320
Lunatia	G	0.5	145	2475	190
Fasciolaria (?)	G	57	1350	950	205
Gervillia	P	1	355	2450	850
Locality 26					
Fasciolaria culbertsoni (?)	G	1	430	1900	440
Sphenodiscus	C	17.5	635	3100	410
Eutrephoceras sp. (Nautiloid)	C	5	1095	..	235
Gervillia	P	4	330	1650	175
D. nicolleti	C	2	1295	4000	1025
L. parvula	P	7	1125	1775	230
Concretion	..	34	16,705	1225	725
Locality 21					
D. conradi	C	1.5	245	2300	150
Sphenodiscus	C	20	2550	1950	> 2000
Gervillia	P	2	220	1625	350
Nucula	G	0.5	100	4300	140
Limopsis	P	18	540	1625	370
Anchura	G	37	1050	3000	2000
Concretion	..	28	13,000	1350	775

TABLE 2. *Continued*

Specie	Class*	Per cent calcite	Mg	Sr	Ba
		Protocardia layer			
Locality 6					
Nodose scaphite (Undescribed)	C	5	270	> 5000	1075
Sphenodiscus	C	8.5	730	3800	1200
Pteria nebrascensis	P	95	1845	1650	120
Concretion	..	23	15,550	2000	825
Locality 26					
D. nicolleti	C	1	390	..	1325
Eutrephoceras	C	65	1615	2800	400
Sphaeriola	P	1	390	> 10,000	1000
Sphenodiscus	C	20	2015	3900	390
Gervillia	P	2	80	1800	400
Concretion	..	33	16,415	1900	1275
		(TIMBER LAKE MEMBER) *Pteria linguiformis* layer			
Locality 35					
D. nicolleti var.	C	9	2000	4900	625
D. cheyennensis	C	10	4700	4100	675
Sphenodiscus	C	4	58	4500	260
S. nebrascensis	C	3.5	1450	3100	800
Concretion	..	13	> 20,000	1675	1275
Pteria linguiformis	P	2	470	1950	575
Tellina scitula	P	1	100	2300	90
Locality 37					
Mactra warrenana	G	7	170	..	..
Goniomya americana	P	21	12,500	..	..
Idonearca shumarda	P	1	190	2600	280
Fasciolaria buccinoides	G	1.5	120	..	950
Protocardia subquadrata	P	3.5	220	..	850
		Oyster layer			
Locality 16					
Oyster	P	99	410	1950	115
Unidentified clam	P	0	24	1100	370

*C = cephalopod, G = gastropod, P = pelecypod

quantities of the oxides of manganese and iron with high-purity calcium carbonate. The analytical precision is similar for all the elements determined—approximately 10 per cent expressed as coefficient of variation. For the calcite determination the coefficient of variation is about 5–10 per cent for intermediate values and increases to 20 per cent at low concentrations. The detectability limit is about 0.5 per cent calcite.

RESULTS

Table 2 gives the results on a wide variety of fossil-shell material from the Fox Hills Formation and the underlying Pierre shale.

Table 3 gives the visual physical observations and associated analytical data for various samples of shell material from a single *Sphenodiscus* shell.

Table 4 gives similar information on a number of samples taken from *Sphenodiscus* specimens from several different concretions, representing a variety of stratigraphic zones and collecting localities in the Fox Hills Formation.

The visual physical observations and associated analytical data indicate that a wide range of values for all the analytical parameters exists both for samples from a single specimen and for a large number of specimens of the same genus. These data combined with the data of

Table 2 indicate that no single observation on a fossil shell guarantees information on the original composition of the shell.

DISCUSSION

General Considerations

Table 2 shows that the assemblage of fossil tests from the Fox Hills Formation is generally higher in magnesium, strontium, and barium microscopic structures with those of a Recent *Nautilus*. The shells in all cases are composed of very small aragonite crystals with elongation perpendicular to the shell surface. Under crossed nicols, extinction occurs parallel and perpendicular to the "fibrous" aragonite crystals. Growth bands marked by layers of organic material parallel to the shell surface are prominent in the Recent shell but more diffuse in the fossil shells. The *Nautilus* showed fine

TABLE 3. ANALYSES OF CHIPS FROM A SINGLE SPECIMEN OF *Sphenodiscus* FROM LOCALITY 30: *P. linguiformis* LAYER

Specimen	Description	Per cent calcite	Trace elements (ppm) Fe	Mn	Mg	Sr	Ba
1A	Dull, chalky, flakes of shell from interior cavity wall. Faint but visible manganese staining present	2.0	250	1100	115	3900	140
1B	Dull, yellowish white, chalky flakes from layer of shell below 1A	0.0	110	480	82	4400	130
1C	This series of samples represents a single	5.5	620	3500	1500	5400	420
1D	large chip broken from the dorsal edge	4.0	200	1500	460	>6400	370
1E	of the shell. All specimens are identical	3.5	230	2200	790	5900	320
1F	in appearance—fresh-looking, translu-	3.0	400	1800	710	>6500	350
1G	cent, brownish shell. Sequentially sam-	2.5	100	1150	370	>6400	400
1H	pled from the outside in	1.5	113	530	160	5800	350
1I	Chalky, white, brittle flakes from weathered portion of shell	0.5	620	380	48	3600	130
1J	Powder scraped from layer of weathered shell below I	0.5	380	480	50	4300	150
1K	Fairly fresh, dark-brown shell from outer edge of unweathered zone	1.0	240	1000	111	5100	240
1M	Massive, fresh-looking brown shell	0.5	190	620	112	5000	230

than recent molluscan shells which contain on the average 200, 1600, and 10 ppm of these elements respectively (Turekian and Armstrong, 1960).

The most striking disparity is in the much higher barium concentration of the fossil shells. If this is all originally deposited barium, then it implies a higher-than-present Ba/Ca ratio in the Cretaceous sea from which the shells were deposited. The alternative is that the barium was added later.

Presumably the variable calcite percentages are due to secondary effects rather than to primary deposition. Modern species of the genus *Tellina* are all completely aragonitic, whereas the range in the Fox Hills is from 1 to 10 per cent calcite. The modern *Nautilus* is also pure aragonite, but the Fox Hills cephalopods all show wide variations, presumedly as the result of secondary processes.

We have made thin sections of a *Sphenodiscus* and a *Discoscaphites* shell and compared their lines of brownish, presumably organic, material perpendicular to the shell surface. The *Sphenodiscus* shell shows this same structure, although much subdued.

In the fossil shells aragonite crystals are somewhat larger than in the modern shells. This could be a species effect or a diagenetic effect. The *Sphenodiscus* shell shows some fracturing but with no obvious alteration along the fractures. We conclude that the observed fossil shells have retained their original structure but have probably lost some of their interstitial organic material during diagenesis.

Sphenodiscus Study

To gain more insight into the changes taking place in fossil shells, we made an intensive study of the distribution of the trace elements in one organism. *Sphenodiscus* was chosen because: (1) it is a characteristic faunal element of the Fox Hills; (2) in a preliminary study it showed a great variability in chemical composition and

in calcite-aragonite ratios; and (3) it is the largest and most easily sampled organism. The data to be discussed are in Tables 3 and 4.

We have arbitrarily used the percentage of calcite of the sampled material as the monitor of diagenetic change. Tables 3 and 4 show that some of the chalkiest material has the lowest calcite content. We avoided sampling obviously secondary calcite crystals associated with the shell material; however it is still possible that some of the variable calcite percentages may result from contamination by secondary calcite in small voids or fractures in the shell structure.

Figure 2 shows the results for strontium, barium, magnesium, iron, and manganese. The data for a single multiply sampled specimen (Table 3) and the results on a large number of different individuals (Table 4) are plotted. Both sets of data appear to conform to the same general pattern of composition variations.

The strontium curve is the most difficult to

TABLE 4. ANALYSES OF CHIPS FROM *Sphenodiscus* SHELLS FROM DIFFERENT LAYERS AND GEOGRAPHIC LOCATIONS

	Locality and layer*	Description	Per cent calcite	Trace elements (ppm) Fe	Mn	Mg	Sr	Ba
2	35 P.L.	Fairly fresh shell flakes from umbilical region	3.0	4300	2800	800	>6200	470
3	35 P.L.	Fresh, light-brown shell flakes from outer rim of shell	1.0	580	480	94	4700	310
4	34 ?	Fresh dark-brown shell flakes from dorsal edge of shell	6.5	590	1600	1500	>7500	1000
5	30 P.L.	Flakes and chunks or dark-brown shell from umbilical region. Some white chalky material on parting surfaces	10.0	1300	3900	1640	4700	200
6	37 P.L.	Somewhat weathered, thin shell flakes. Sample not too clean—visible staining with iron and manganese. Flakes taken from near, but not on, the dorsal edge	>36.5	600	>4300	7500	3900	2200
7	37 P.L.	Fairly fresh flakes from side of shell. Two distinct but inseparable layers present in sample; outer layer light brown, inner layer somewhat darker	12.0	3300	3400	3900	5600	500
8	25 L-G	Flakes from side of shell. Fresh looking but locally recrystallized; color, yellowish white	>65.5	5800	>7000	10100	2100	520
9	4 P	Massive, very dark-brown chunks of fairly fresh shell from umbilical region	2.5	960	2800	830	≫8000	1200
10	8 L-G	Pure crystalline calcite filling and shell. Sample almost entirely calcite which is clean, yellowish	95.0	11000	3900	8200	900	300
11	8 L-G	Same specimen as 10. Small flakes of pure shell. Flakes all chalky white and iridescent	8.5	2500	2800	1360	4200	2500
12	W	Weathered, whitish, manganese-stained flakes of chalky shell material	0.0	580	840	105	4900	130
13a	25 L-G	Iridescent whitish shell material. Each sample represents a separate individual	18.0	1500	2000	4500	3300	580
13b			0.0	400	390	210	3400	160
14	30 P.L.	Flakes and chunks of dark-brown, fresh-looking shell from umbilical region and side of the animal	3.5	3100	1500	1300	>7300	880

*L-G = *Limopsis-Gervillia;* P = *Protocardia;* P.L. = *P. linguiformis;* W = Specimen from Wyoming, unknown layer

understand, considering past reasoning on the direction of diagenetic effects on this element in calcareous materials. The highest strontium values are between 2 per cent and 10 per cent calcite, and these are all much higher than modern molluscan shells. This cannot be due to incomplete separation from the matrix, since Table 2 shows that the concretionary material has about the same (or in some cases even lower) barium concentration as the shells.

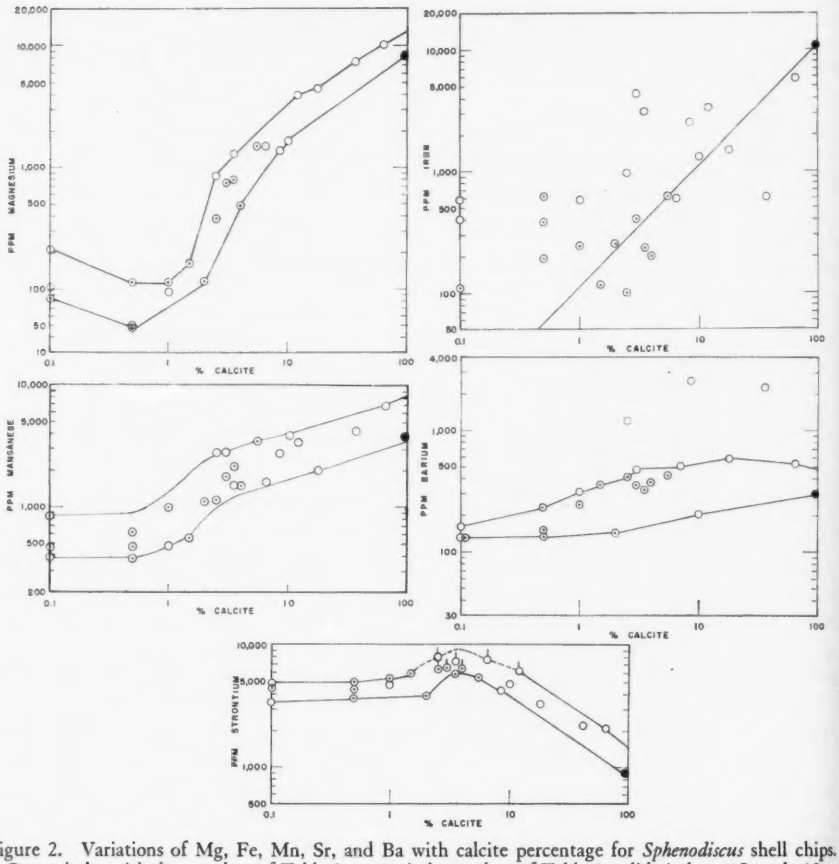

Figure 2. Variations of Mg, Fe, Mn, Sr, and Ba with calcite percentage for *Sphenodiscus* shell chips. Open circles with dots = data of Table 3; open circles = data of Table 4; solid circles = Sample 10

contemporary molluscan tests (average 1600 ppm Sr).

In the case of barium the trend indicates a slightly increasing barium concentration with increase in per cent calcite. The three very high values may be due to the presence of barite. However, even the almost pure aragonite shells show an average of about 130 ppm Ba, more than 10 times the average for

The magnesium concentration shows an increase of a factor of 100 between the aragonitic and calcitic end members of the series. The manganese also shows a marked increase. The low-calcite values of about 500 ppm Mn are 100 times higher than those of contemporary molluscan shells (Krinsley, 1959). The iron shows an increase with increasing calcite content, but the spread of values is larger. Even

TABLE 5. LEACH EXPERIMENT DATA ON SINGLE *Sphenodiscus* SHELL CHIP
Each leach fraction represents about 25 per cent of the shell.

Specimen no. 14	Trace elements in parts per million				
	Fe	Mn	Mg	Sr	Ba
L1	2750	1900	1000	>7000	900
L2	2800	1800	1100	>7000	800
L3	2000	1300	950	>7000	900
L4	1500	1100	1000	>7000	800
Residue	Very high				Very high

the nearly pure aragonite samples are very much higher in iron than contemporary molluscan shells.

To test the possibility that all the high values for the trace elements and the high calcite values were superficial we subjected a single large piece of *Sphenodiscus* shell to a series of acid leaches. The dissolved calcium carbonate and trace elements from each leach were reprecipitated and analyzed in the same manner as the shell material.

Each successive leach sample represented about 25 per cent of the shell material. Table 5 shows the results. Iron and manganese definitely decrease in the later fractions. Within the analytical errors magnesium, strontium, and barium appear to be evenly distributed through the shell, although the next to the last fraction might have higher strontium than the others. In any case the values are all very high. A residue that did not dissolve in hydrochloric acid was very high in barium and iron, suggesting the presence of small amounts of barite, iron oxides, and clay or glauconite.

We also leached a number of other specimens with acid. But in the experiments reported in Table 6 roughly 50 per cent of the shell material was dissolved. Analytical data are presented for the leached fraction, precipitated as before, and the residual fraction, probably containing acid-insoluble fractions in some cases. The composition close to that of the original shell can be gotten from the corresponding numbers of Table 3 and 4, but because of the heterogeneity observed even in small parts of the shell (See specimens 1C to 1I, Table 3) the correlation between the average of the residue and leach values and the corresponding values for the original shell is not always good.

Significance

We conclude that a single analysis of shell material as given in Table 2 does not necessarily represent the composition of the shell as it occurs in the concretion. Neither does a set of analyses, although they may show a trend, guarantee that the samples with the lowest calcite percentages are representative of original shell material.

We have two options of interpretation: either the lowest-calcite-percentage shells are the closest approximation to the original shell, or the entire complexion of analyses represents

TABLE 6. ACID LEACH DATA FOR VARIOUS *Sphenodiscus* SHELL CHIPS
Each sample was leached with an amount of hydrochloric acid to dissolve 50 per cent of the calcium carbonate. The leached fraction is designated L and the residue R.

Specimen	Per cent calcite	Trace elements in parts per million				
		Fe	Mn	Mg	Sr	Ba
3R	0.0	250	190	61	5300	240
3L		290	780	100	4700	250
5R	2.0	330	1500	520	4800	200
5L		270	>4200	1500	4700	200
7R	21.0	5400	>5300	8400	4900	500
7L		1300	3500	4200	5700	500
9R	5.5	970	3900	1220	≫8000	1200
9L		900	>4200	1700	≫8000	1200
14R	0.0	1200	1050	330	>6900	700
14L		760	1600	520	>6200	700

multiple effects of diagenesis and weathering.

The first option loses some of its attractiveness because of the inordinately high iron and manganese concentrations compared to those of modern shells. Although undoubtedly much of this is surficial, as evidenced by the leach experiments, enough of it has become so closely associated with the calcium carbonate that it is inseparable from it.

If we reject the data on Mn and Fe because these are secondary and consider the data on barium and strontium, we find that the strontium concentration is still higher than that of most modern molluscan shells by a factor of 2 (when averages are concerned), whereas the barium concentration is about 10 times higher. If this is related to the barium-calcium ratio of the surrounding sea water from which the animal deposited its shell, then the Fox Hills sea water must have had about 10 times higher Ba–Ca ratio than contemporary shelf ocean water.

On the other hand, if we assume that the range of data on *Sphenodiscus* is the result of diagenetic processes, the variability and the high values of the trace-element concentrations at low per cent calcite are a little more comprehensible.

We have attempted to construct a quantitative model of the effects of different types of diagenetic change we think might shed light on our data. Our numerical results *per se* have dubious value, but they illustrate the direction of thinking that our studies suggest.

Let us assume that in each of the *Sphenodiscus* shell chips we have analyzed there are three distinct phases: (1) original unaltered shell material composed of pure aragonite; (2) completely recrystallized material, either replacing the original shell or filling interstices, composed of pure calcite; and (3) a "reaction layer" in which there is adsorbed material on otherwise unaltered aragonitic shell material. This last may be the consequence of about 3 per cent organic material (Thompson and Chow, 1955) as a network within the shell, which undergoes decomposition and leaves both vacant interstices and perhaps a reactive environment for adsorption processes on the organic decay fragments.

We assume that these alterations described in (2) and (3) proceed at different rates in different environments, hence the ratios of the three phases are not simply predictable from the calcite concentration. In this model, however, 100 per cent calcite is taken to mean the absence of both the original shell and the "reaction layer".

Each of the three phases is assumed further to have a distinctive trace-element composition. This is certainly one of the poorest assumptions, since not only are there marked variations in original shell composition within a genus, but also the composition of the two altered phases must be a function of the solutions to which a particular shell fragment is subject. We have ignored these variations in the framing of the model. The size of the errors on our computed values is undoubtedly a reflection of this as well as the naivete of the three-phase model itself.

Any particular shell chip then has a concentration of trace elements reflecting the different fraction of each phase and its distinctive contribution of trace elements. That is:

$$C_{im} = a_m x_i + b_m y_i + c_m z_i$$

where:

C_{im} = measured concentration of element m in shell chip i

x_i = fraction of original aragonitic unaltered material in shell chip i

y_i = fraction of calcite in shell chip i

z_i = fraction of "reaction layer" aragonitic in composition in shell chip i

and

$x_i + y_i + z_i = 1$

a_m, b_m, c_m = the concentrations of the trace element in the phases x_i, y_i, and z_i respectively.

To determine the fraction of each phase in each shell chip it was necessary to use the strontium data, since only strontium goes through a maximum in Figure 2. We chose a_{Sr} = 2500 ppm, b_{Sr} = 1000 ppm, and c_{Sr} = 10,000 ppm. The original shell value (a_{Sr}) is the average of values that we, Odum (1957), and Thompson and Chow (1955) found for *Nautilus*. The calcite value (b_{Sr}) obtained from the analysis of a calcite crystal growing in one of the cavities in a *Sphenodiscus* is shown in Figure 2 and Table 4 (sample 10). The value of c_{Sr} is an assumed value based on the high values of the Sr curve in Figure 2. The value cannot be lower than about 8000 ppm, and although we cannot reject a value higher than 10,000 ppm we have preferred a low value. We are aware that the choice of the value of c_{Sr} seriously influences the results of our computation. Table 7 gives values of x, y, z for all specimens calculated on these assumptions.

TABLE 7. ESTIMATED PROPORTIONS OF ORIGINAL SHELL AND ALTERATION PHASES CALCULATED FROM THE STRONTIUM DATA

Assuming: a_{Sr} = 2500 ppm, b_{Sr} = 1000 ppm, c_{Sr} = 10,000 ppm. (See text.) x = original shell; y = calcite; z = "reaction product"; x + y + z = 1

Specimen	x	y	z
1A	0.77	0.02	0.21
1B	0.75	0.0	0.25
1C	0.55	0.055	0.395
1D	0.43	0.04	0.53
1E	0.505	0.035	0.46
1F	0.43	0.03	0.54
1G	0.45	0.025	0.525
1H	0.54	0.015	0.445
1I	0.85	0.005	0.145
1J	0.75	0.005	0.245
1K	0.64	0.01	0.35
1M	0.66	0.005	0.335
2	0.47	0.03	0.50
3	0.70	0.01	0.29
4	0.26	0.065	0.675
5	0.59	0.1	0.31
6	0.38	0.365	0.255
7	0.44	0.12	0.44
8	0.265	0.655	0.08
9	0.235	0.025	0.74
11	0.67	0.085	0.255
12	0.68	0.0	0.32
13a	0.68	0.18	0.14
13b	0.88	0.0	0.12
14	0.32	0.035	0.645

Using the values of the concentrations of each of the other trace elements for each shell chip with the corresponding values of *x*, *y*, and *z* calculated from the Sr data, we obtained the best values of *a*, *b*, and *c* for each element by solving all 25 equations using a high-speed computer (IBM 610). Table 8 gives these values with reasonable values for modern molluscan shells for comparison.

The results are not spectacular, which is not surprising considering the large number of assumptions. However, the model serves as a device to encourage the viewing of a fossil shell of any age as a complex of alteration products. It is important to understand the sequence and rates of these alterations. Perhaps a study of the uranium decay series in shells of known age from a variety of sedimentary environs will give us some clues.

CONCLUSIONS

(1) Visual appearance of shell material is not necessarily a valid criterion of degree of alteration.

(2) Even where there does not appear to be any alteration in mineralogy we have found an alteration of the original chemical composition of the shell. We suspect that this chemical alteration occurs in that part of the shell originally occupied by organic material.

(3) The results presented here cast doubt on the use of the composition of molluscan external hard parts for studies in paleoecology. Studies on materials that might be of potential use in radioactive dating must be undertaken with caution.

TABLE 8. CALCULATED COMPOSITIONS OF ORIGINAL *Sphenodiscus* SHELL AND ALTERATION PHASES USING DATA OF TABLES 3, 4, AND 7

In parts per million; a, calculated concentrations in original shell; r_a, standard deviation on a; b, calculated concentration in calcite; r_b, standard deviation on b; c, calculated concentration in "reaction layer"; r_c, standard deviation on c; A estimated values on modern shells; B, concentration in pure calcite chip from shell cavity.

	a	r_a	b	r_b	c	r_c	A	B
Fe	114	235	7173	1418	1485	471	$\lesssim 10$	11,000
Mn	278	158	10,635	954	2645	317	$\lesssim 10$	3900
Mg	−21	120	17,101	724	640	240	$\lesssim 100$	8200
Ba	52	113	1905	681	1035	226	~15	300

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Short Notes

J. PHILIP SHANNON, JR.

UPPER PALEOZOIC STRATIGRAPHY OF EAST-CENTRAL IDAHO

Abstract: Mississippian, Pennsylvanian, and Permian rocks exposed in the Lemhi and Lost River ranges of east-central Idaho comprise a conformable succession of limestone and sandstone units. These strata have been correlated by earlier workers with the Mississippian Brazer Limestone of northern Utah and have been referred to by that name. Since the east-central Idaho section bears little lithologic or temporal similarity to the Brazer, the name should be abandoned in that area and new names applied to units of proved mappability.

Introduction

Previous workers have referred more than 5000 feet of strata in east-central Idaho to the "Brazer" Limestone, a name applied to Upper Mississippian limestone in northern Utah and southeastern Idaho (Richardson, 1913; Mansfield, 1927). Umpleby (1917) and Ross (1934b; 1937; 1947) considered "Brazer" Limestone north of the Snake River downwarp to be Mississippian on the basis of time-stratigraphic correlation with the lithologically dissimilar Upper Mississippian Brazer Limestone at its type locality to the southeast. Blackstone (1954) discovered Permian fusulinids in limestone at the southern tip of the Lemhi Range. This has suggested the presence of a complete section of Upper Paleozoic rocks in the so-called Brazer of east-central Idaho. The writer discusses the Upper Paleozoic stratigraphy of two areas in the southern Lemhi and Lost River ranges (Fig. 1).

Upper Paleozoic strata are exposed along the west flank of the southern Lemhi Range and on the east and west sides of the southern extension of the Lost River Range in the Howe Peak area. Relief in the areas exceeds 1 mile (elevation of the valley floor is 5200 feet) and results from post-Laramide block faulting that produced several subparallel northwest-trending ranges in east-central Idaho; intense Laramide compression preceded the block faulting. The stratigraphic columns in Figure 2 are composites of sections measured along the western faces of the ranges, except for the upper portions of the Howe Peak section, which were measured on the east flank of the range. Here the youngest strata are exposed in a series of bevelled anticlines and synclines.

Acknowledgments

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Stratigraphy

The Upper Paleozoic section includes five distinct lithologic units (Fig. 2: "A", "B", "C", "D", "E", from oldest to youngest). The mappability of these units and their status as potential formal rock units has been demonstrated by maps of the southern Lemhi Range (J. F. King, 1959, M. S. thesis, Northwestern Univ.) and Howe Peak area (J. P. Shannon, Jr., 1960, M. S. thesis, Northwestern Univ.), on which the several units have been traced over an area of 150 square miles. Outside these two areas units "A", "B", "C", and "D" have been recognized in the vicinity of Arco Peak, 18 miles to the west near the town of Arco, Idaho (index map, Fig. 1). Further extent of these subdivisions is unknown, although thick sections of "Brazer" Limestone in the Borah Peak area to the northwest (Ross, 1947, p. 1116) and the Beaverhead Range to the east (Scholten, 1957a, p. 164) offer the possibility of broadening the areal limits of recognition of the units.

Unit "A" is medium-gray calcareous sandstone, sandy limestone, and olive-green to light-gray quartzose sandstone, all of which

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weather to form a characteristic smooth brown slope. The unit is conformable with an underlying dark-purplish-gray flaggy shale, considered by the writer to be the eastern equivalent of the Milligen Formation (described in the Wood River area by Umpleby, Westgate, and Ross, 1930, p. 25). Unit "A" is 775–820 feet thick and shows a notable increase in carbonate content toward its top. A distinctive feature is the presence of massive dark-gray lenticular limestone lenses which attain thicknesses of 50–60 feet. Although these have a maximum observed lateral extent of about 3000 feet, their unique occurrence in unit "A" provides for easy recognition and certain traceability of that unit.

No diagnostic fossils have been found in unit "A", but conformable relationships with the Milligen Formation below, in part Lower Mississippian (Sloss, 1954, p. 368), and the Upper Mississippian unit "B" above indicate that the intervening section is probably Osagean and/or Meramecian.

Unit "B" comprises 1350–1550 feet of dark-gray fossil-bearing limestone that characteristically forms imposing cliffs. Massive bedding, dark color, and abundant large horn corals distinguish this unit from other limestones in the stratigraphic section. Numerous sandy zones are present throughout the unit, as are nodules, lentils, and stringers of black chert. The large horn corals *Caninia*, *Dibunophyllum*, and *Faberophyllum*, and the Foraminifera *Millerella* and *Paramillerella* suggest that unit "B" is Meramecian, Chesterian, and possibly Early Pennsylvanian. Other fossils, *Dictyoclostus*, *Linoproductus*, *Syringopora*, and *Lithostrotionella* confirm the Mississippian and/or Pennsylvanian age.

Unit "C", like unit "A", weathers to a smooth brown slope. It crops out as a narrow band 190–230 feet thick above the massive dark limestone of unit "B" and is conformable with the lower unit. The unit is composed primarily of relatively pure, white, light-gray, and olive-green quartz sandstone, with thin interbeds of light-gray limestone. This lithology, unique in the Upper Paleozoic section, along with the characteristic weathering, makes this unit an excellent mappable marker zone. Since unit "C" is unfossiliferous, its age can only be inferred from the adjacent strata. The lower portion of the conformable younger unit "D" is Atokan; the upper portion of unit "B" is considered to be as young as Early Pennsylvanian. The writer thus tentatively assigns unit "C" to the Atokan.

Unit "D", about 3000 feet thick, crops out over the entire eastern half of the Howe Peak area and forms the southern toe of the Lemhi Range. The unit is a light- to medium-gray limestone containing numerous bands, lentils, and nodules of dark-gray and black chert; oölitic zones; and large concentric chert structures up to 6 inches in diameter. The large number of cherty zones imposes some difficulty in using any one as a marker zone. Unit "D" is lighter in color, thinner-bedded, and more shaly than the massive-bedded unit "B" and is less resistant to weathering and erosion. These physical characteristics help to distinguish the two units in the field. The writer did not subdivide the unit because of the difficulty in recognizing distinctive or traceable lithologic zones.

Among the megafossils *Neospirifer* indicates a post-Mississippian age. Also present, but less diagnostic, are *Dictyoclostus* and the corals *Syringopora*, *Lithostrotionella*, and *Tryplophyllites*. Sandy and cherty zones throughout the unit contain fusulinids (Fig. 2). Unfortunately, these zones are not lithologically distinctive, and consequently the writer could not trace them from one area of study to the other. Recognition of the Atokan form *Fusulinella* low in the unit and a fusulinid similar to the Leonardian form *Parafusulina* near its top indicate that unit "D" ranges in age from Atokan to and possibly including Leonardian. *Fusulina megista* (Desmoinesian) and *Triticites*, *Oketaella*, *Triticites cellamagnus*, and *Schwagerina* (transitional Upper Pennsylvanian-Lower Permian) suggest the presence of Desmoinesian, Missourian, Virgilian, and Wolfcampian strata.

The Atokan Series in both areas is relatively thick; however, beds of Missourian and Virgilian age appear to be considerably thinner in the Lemhi Range than in the Howe Peak area. Dating of the Pennsylvanian portion of the section is still too scanty to establish series boundaries with certainty.

Unit "E" comprises at least 200 feet of light-gray quartz sandstone, dolomitic sandstone, and sandy limestone that form soft grassy slopes on the eastern flanks of the ranges. These slopes reveal a sparse scattering of outcrops and grade imperceptibly into large alluvial fans near the bases of the ranges. Owing to the poor exposures the writer has not been able to determine either thickness or lithology of the unit satisfactorily. As Figure 1 shows, the unit is probably Leonardian, since it conformably overlies unit "D", which is Wolfcampian, or Leonardian at youngest.

Figure 1. Index map and correlation of formations, Montana, Idaho, and Utah

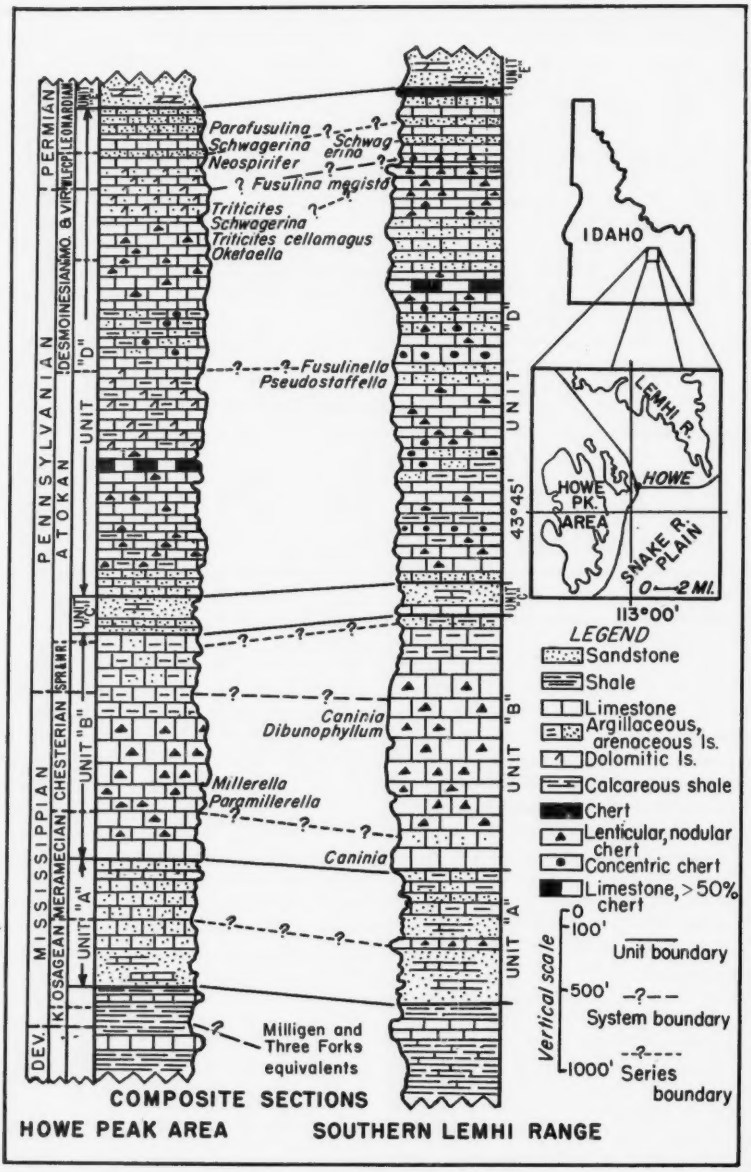

Figure 2. Composite stratigraphic sections, Howe Peak area and Southern Lemhi Range, Idaho

Regional Relationships

The writer differentiates unit "A" from the Milligen Formation on the basis that only the lower dark-gray shale resembles the Milligen Formation at its type locality to the west, where that formation consists of black argillite and is virtually free of any calcareous material. The carbonate content and lensing limestones above the Milligen Formation warrant the establishing of a separate unit whose basal portion is possibly the time equivalent of part of the Milligen Formation to the west in Muldoon Canyon and the Wood River valley. M. R. Thomasson (1959, Ph.D. thesis, Wisconsin Univ.) has described 8500 feet of highly conglomeratic sandstone and shale with interbedded limestone in Muldoon Canyon (Fig. 1) that occupy a stratigraphic position between the Milligen and Wood River formations. The writer considers the upper portion of unit "A" and the major part of unit "B" to be the time equivalents of that unit (Fig. 1).

The Mississippian System is represented to the northeast by the Madison and Big Snowy groups and lower Amsden Formation (Sloss and Moritz, 1951, p. 2161) and to the southeast by the Madison Group and Brazer Formation. Inasmuch as the Upper Paleozoic succession in east-central Idaho is only in part the lithologic and temporal equivalent of the Brazer Formation south of the Snake River Plain, the writer suggests that the formation name be restricted to the section in southeastern Idaho and northern Utah. In fact, only about 1500 feet of the east-central Idaho section can be equated in time with the Brazer Formation; the majority of that succession has time equivalents in the Amsden, Quadrant, and Phosphoria formations of southwestern Montana (Scholten, 1957b, p. 1794) and the Morgan, Weber (Wells), and Phosphoria formations to the southeast (Fig. 1).

M. R. Thomasson (Personal communication, 1959) has suggested that unit "C" might be an eastern extension of the basal conglomeratic unit of the Wood River Formation. However, the clean nature of the sand, the thinning of the unit from east to west (Fig. 2), and the resemblance to cratonically derived detritus indicates that this unit is more likely a western extension of the Quadrant Formation. Nevertheless, the Atokan age of the unit fits well with either interpretation. Unit "D" is definitely the time equivalent of the Wood River Formation in the Wood River valley (index map, Fig. 1). Bostwick (1955) has recovered fusulinids from the section in that area that establish the age of the Wood River Formation as Middle and Late Pennsylvanian and Early Permian, an age which agrees with that of unit "D". However, unit "D" is obviously not the lithologic equivalent of the Wood River Formation, a series of sandy limestones and calcareous sandstones, and consequently should not be referred to by that name.

No time equivalent of unit "E" has been recognized in the Wood River valley; furthermore, rocks similar to this unit in the intervening regions have been reported in only the Hailey quadrangle (Ross, 1934a, p. 990), so present consideration of unit "E" as an eastern extension of the Wood River Formation is unwarranted. McKelvey *et al.* (1956, p. 2840–2846) have subdivided the Permian rocks of southeast Idaho into the Park City (Upper Pennsylvanian-Lower Permian) and Phosphoria (Middle Permian) formations; unit "E" is probably the time equivalent of the lower portion of the Phosphoria Formation.

The writer has shown that Upper Paleozoic rocks in east-central Idaho comprise the Osagean or Meramecian through Leonardian Series and include several gradational but lithologically distinct units. These units bear little lithologic or temporal relationship to the Upper Mississippian Brazer Formation of southeastern Idaho and northern Utah. The term "Brazer" can no longer be considered valid in east-central Idaho and should be discontinued when new names are formally assigned to units of proved mappability.

Stratigraphic Sections

Southern Lemhi Range, Butte County, Idaho

Secs. 16, 21, T.6 N., R.30 E.	Thickness (feet)
Top of exposures	
Unit "E"	
Interbedded calcareous and dolomitic quartzose sandstone and sandy limestone	200+
Unit "E" (minimum thickness)	200
Unit "D"	
Chert, thick-bedded, medium gray, weathers brown to red	20
Limestone, light gray to medium gray, finely crystalline, weathers light gray to buff; interbeds of light-gray quartz sandstone; *Lithostrotionella* scattered throughout unit	359
Limestone, light gray to medium gray,	

	Thickness (feet)
finely crystalline; contains *Lithostrotionella*, *Schwagerina*	14
Limestone, light gray to pink, finely crystalline, weathers light gray to buff; shaly zones, as well as zones of black nodular chert	165
Limestone, dark gray, finely crystalline, with scattered chert nodules; *Fusulina megista*	10
Limestone, light gray to medium gray, finely crystalline, with scattered small chert nodules and concentric chert; *Lithostrotionella*, *Dictyoclostus*, gastropod fragments	208
Quartzose sandstone, light gray to medium gray, calcareous, weathers light gray to buff; interbeds of light- to medium-gray, finely crystalline, sandy limestone	335
Limestone, medium gray, medium to coarsely crystalline, weathers medium gray; zones composed almost entirely of black-chert nodules; bryozoans and other fossil fragments scattered throughout	200
Limestone, medium to dark gray, finely crystalline, weathers to light-gray smooth slope; bryozoans scattered throughout unit, but especially profuse near top; black nodular and concentric chert; middle portion of unit sandy	260
Limestone, dark gray, finely crystalline, weathers medium gray; concentric chert, *Tryplophyllites*, *Pseudostaffella*, and *Fusulinella* present near center of unit	102
Limestone, medium gray to dark gray, alternating oölitic and fossil fragmental, weathers light to medium gray; zones of black lenticular chert; *Syringopora;* sandy near top of unit	438
Limestone, light gray to buff, fine to moderately crystalline, weathers light gray; sandy and shaly zones in lower half, black-chert nodules and concentric chert in upper portion	289
Limestone, medium gray to dark gray, finely crystalline, weathers medium gray; shaly zones scattered throughout unit; black-chert nodules in center and upper portions, large concentric chert near base; oölitic zones in upper 35 feet	314
Limestone, medium gray, finely crystalline, weathers light gray; nodules and stringers of black chert	144
Limestone, medium gray to dark gray, finely crystalline, weathers brown to gray buff; nodular black chert; sandy near base of unit	155
Unit "D"	3013

	Thickness (feet)
Sec. 3, T.6 N., R.29 E.	
Unit "C"	
Quartz sandstone, medium gray to light gray, fine-grained, weathers medium brown; limestone concretions	45
Quartz sandstone, medium gray to light brown, light green, fine-grained, weathers brown to olive green; interbeds of limestone, silty, medium gray, finely crystalline, weathers light gray to buff; forms smooth brown slope	105
Limestone, dolomitic, light gray to buff to pink, finely crystalline, weathers light gray to buff; interbeds of light-gray quartz sandstone near base of unit; forms smooth brown slope	40
Unit "C"	190
Sec. 6, T.6 N., R.30 E.	
Unit "B"	
Limestone, medium gray to dark gray, finely crystalline, weathers dark gray; shaly zones in lower portion; *Syringopora*, *Caninia*	266
Limestone, dark gray, dense to finely crystalline, weathers dark gray; scattered black-chert nodules and stringers; *Dibunophyllum*, *Faberophyllum*, *Gigantella* near base	530
Limestone, dark gray to black, massive, finely crystalline, weathers light gray to medium gray; black-chert nodules present, also scattered large horn corals	468
Sec. 1, T.6 N., R.29 E.	
Limestone, dark gray to black, thin-bedded, finely crystalline to fragmental, weathers dark gray; *Tryplophyllites*, *Caninia;* numerous oölitic zones	150
Limestone, medium gray, light gray near base, massive, moderately crystalline to coarsely fragmental; dark gray to black chert scattered throughout unit	130
Unit "B"	1544
Unit "A"	
Limestone, sandy and shaly, dark gray, thin-bedded, finely crystalline, weathers rust brown; forms smooth brown slope	270
Quartz sandstone, calcareous, fine-grained, weathers rust brown; forms smooth slope	135
Limestone, dark gray, finely crystalline; black-chert stringers in 1–2-inch bands; forms prominent ledge	66
Quartz sandstone, calcareous, medium gray, fine-grained, weathers olive green to rust brown; interbeds of limestone, medium gray to dark gray, finely crys-	

	Thickness (feet)
talline; forms smooth slope broken by limestone ledges	337
Unit "A"	808
Total thickness	5756

Howe Peak area, southern Lost River Range, Butte County, Idaho
Sec. 4, T.5 N., R.29 E.

Top of exposures	
Unit "E"	
Interbedded dolomitic and calcareous quartzose sandstone and sandy limestone	220+
Unit "E" (minimum thickness)	220
Unit "D"	
Limestone, silty, light gray to medium gray, finely crystalline, with sandstone stringers	157
Limestone, light gray to medium gray, finely crystalline, with *Dictyoclostus, Schwagerina, Parafusulina*	20
Limestone, light gray to medium gray, fine to moderately crystalline, weathers medium gray to light brown; contains several sandy zones and zones of calcareous sandstone; *Lithostrotionella, Neospirifer*	182
Limestone, silty, dolomitic, medium gray to light gray, finely crystalline to saccharoidal; black-chert lenses and nodules scattered throughout	304
Sec. 1, T.4 N., R.28 E.	
Limestone, medium gray to dark gray, finely crystalline, weathers medium gray; contains *Triticites cellamagnus, Schwagerina, Oketaella*	33
Limestone, medium gray to light gray to buff, finely crystalline, with interbeds of sandstone showing excellent cross-bedding; light- to medium-gray chert nodules scattered throughout, also black lenticular chert and concentric chert; *Lithostrotionella*	901
Limestone, medium gray, fine to moderately crystalline, with dolomitic and shaly zones	479
Limestone, medium to dark gray, thin-bedded, finely crystalline, weathers medium gray; brown and black cherty zones consisting of 60–70 per cent chert; bryozoan fragments, *Syringopora*	316
Limestone, silty to sandy, medium gray, finely crystalline, weathers light gray to buff	164
Limestone, medium gray to dark gray, finely crystalline, weathers medium gray to brown; scattered shaly zones; brown and black chert nodules and stringers throughout unit	232
Limestone, medium gray, moderately crystalline to coarsely fragmental, weathers gray to brown; interbeds of brown quartz sandstone; productids scattered throughout	153
Unit "D"	2941
Sec. 22, T.4 N., R.28 E.	
Unit "C"	
Quartz sandstone, medium gray to light gray, fine-grained, weathers medium brown; limestone concretions	35
Quartz sandstone, light gray to blue green, fine- to medium-grained, weathers light brown to olive green; interbeds of medium-gray silty limestone; forms smooth brown slope	118
Limestone, sandy, medium gray to light gray, finely crystalline, weathers brown	45
Quartz sandstone, calcareous, buff, fine-grained, weathers brown; bedded light-gray chert	30
Unit "C"	228
Unit "B"	
Limestone, silty to shaly, medium gray, finely crystalline, weathers light gray to buff; sandstone stringers throughout	297
Limestone, shaly, medium gray, finely crystalline, weathers brown to medium gray; lenticular medium-gray to black chert	185
Limestone, massive, dark gray, finely crystalline, weathers dark gray to black; black lenticular and nodular chert; *Linoproductus, Dictyoclostus*, large horn corals	415
Limestone, dark gray to black, finely crystalline, weathers black; lenticular gray chert; *Caninia, Millerella, Paramillerella* near top	165
Sec. 8, T.5 N., R.28 E.	
Limestone, medium gray to dark gray, finely crystalline, weathers light gray; black-chert stringers	167
Limestone, medium gray to light gray, massive, moderately crystalline to coarsely fragmental; forms prominent ledge	116
Unit "B"	1345
Unit "A"	
Limestone, light gray to medium gray, moderately crystalline, with fossil fragmental zones; weathers medium gray; sandy near top of unit, with upper 45 feet weathering to a smooth slope	110
Limestone, very silty and sandy, medi-	

	Thickness (feet)		Thickness (feet)
um gray, finely crystalline, weathers brown; forms smooth brown slope	421	shaly zones throughout; forms smooth brown slope	243
Quartz sandstone, calcareous, medium gray, fine-grained, weathers brown;		Unit "A"	774
		Total thickness	5508

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ANCIENT METEORITIC DUST

Abstract: Dominantly magnetic, spherical, and spheroidal microscopic particles (50–850μ) observed in well cuttings and cores of sedimentary rocks of Cretaceous, Miocene, and Pleistocene age from the Sacramento, San Joaquin, and Ventura basins of California respectively, are similar, when particulate in drill cuttings, to weld spatter, material not unexpected at a well site. It is concluded that much of the California material is meteoritic in origin and dates back at least to the Late Cretaceous.

During routine examination with binocular microscope of residues of cores and cuttings of sedimentary rocks analyzed as an aid to stratigraphic correlation, rare, microscopic, highly irregular, dominantly magnetic, dominantly metallic debris was observed and initially regarded as fragments resulting from the wear of drilling bits and other metallic components during rock penetration. This material was found in rock cuttings as old as Cretaceous and as young as Pleistocene from the Sacramento, San Joaquin, and Ventura basins of California. All ages were established by analyses of foraminiferal assemblages.

The widespread horizontal and vertical occurrences of this debris appeared to substantiate the early hypothesis. However, soon after these studies began, very rare, microscopic, spherical, magnetic bodies were observed in the metallic debris. A literature search and a comparison with all material likely to be contributing sources led to the belief that these spheres and spheroids were probably contaminants entering the mud pit and open ditch and were being circulated through the system with the drilling mud.

The initial literature search suggested three probable sources: meteoritic dust, weld spatter (Thomsen, 1952, p. 303–306), or fly ash (Handy and Davidson, 1953, p. 373–379). Metallic grinding dust was briefly considered at a later date (Dr. Claussen, Arcrods Corp., personal communication).

To fix the origin properly, samples of weld spatter were collected for comparison; study of fly ash and meteoritic dust was restricted to scrutiny of photographs and the literature. A sphere was observed *in situ* in a core of Sisquoc claystone of late Miocene age recovered from the offshore portion of the Ventura basin. Other rare spheres were found in residues prepared from other Sisquoc cores. Contamination during laboratory preparation was ruled out so that, when the residues contained spheres, the only conclusion was that the objects had been in the rock material and were at least as old as late Miocene.

These finds were followed by the appearance of the spheres and spheroids in rock chips that had survived the drilling and laboratory processes with their lithology intact (Pl. 2, figs. 2–7). Since 1953 when the first spheres were observed there have been scattered occurrences of these spheres *in situ* in rock chips. When one considers the rigorous treatment to which the rocks are subjected during drilling and processing, and the initial rarity of the spheres, it is surprising that they are found *in situ*.

The basic function of the methods of study, achieved by physical and/or chemical means, is the release and concentration of the diagnostic material from the host rock, enabling the observer to examine most expeditiously the material under investigation. Following wet disaggregation the slurry is washed through a nest of successively finer meshed sieves, the smallest of which will permit the passage of particles one-sixteenth of a millimeter and smaller. The sieve fractions are collected, dried, and examined.

The final residues are spread thinly on a small plastic tray and scanned with a small magnet shielded by a sheet of ordinary paper preventing metal-to-metal contact. The magnet, paper shield, and magnetic debris are moved over another tray, the magnet is removed, and the magnetic material no longer held by magnetic attraction drops. This procedure is the simplest, cleanest, and a positive means of removing microscopic magnetic material. However, since some of the spheres were

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nonmagnetic, scanning was completed with binocular microscope, and removal was with the moistened tip of a number 00 sable hair brush.

Semipermanent mounts are made by coating a cardboard slide, the type used for microfossils, with gum tragacanth. The quick-drying, water-soluble nature of this adhesive makes it desirable, for when a small object is placed on a gum-coated slide with a moistened brush the gum will immediately become soluble, and the object will settle into the viscous layer. Upon drying the object is permanently mounted until removal with a moistened brush. Objects can be reoriented at any time.

had a high rate of sediment deposition. Individuals of all groups (Table 1) have been observed *in situ* in chips.

Almost all the material in groups 1–3 was attracted to a magnetic needle; the attraction ranged from very weak to very strong. Generally the black, nonmetallic individuals in group 1 were in the weaker range. Multiple individuals of two types were observed in groups 1–3: those made up of equal-sized components (Pl. 1, fig. 1) and those of a large sphere upon which were one or more subsidiary spheres (Pl. 1, fig. 2). Most of the spheres sectioned were hollow; wall thicknesses ranged from a few microns to those that were

Table 1. Classification of Spherical Material

Group	Color	Exterior	Magnetic attraction	Wall	Core
1	Black	Smooth, faceted, dull, shiny	Weak to strong	Thick	Yes
2	Steel gray	Smooth, burnished, reticulate	Strong	Thin	None
3	Mottled black, steel gray	Faceted, intergrown	Strong	Thick	None
4	Brown to black	Smooth, rare bubbles	Nonmagnetic	Thick	None

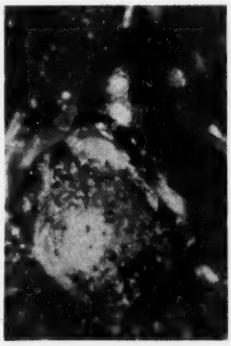
All the individuals figured were photographed using a compound microscope with a 7.8 ocular and a 5-power objective. This system was coupled to a Leitz MIKAS. In most instances depth of focus was materially increased by using the diaphragm procedure presented by Fournier (1956). Three bulls-eye microscope lamps placed at 120-degree intervals around the stage provided illumination.

All sectioning was done on an Arkansas novaculite hone. The spheres were oriented initially by cementing with gum-tragacanth to a glass slide. Permanent mounting was with a droplet of "glue-bird". This material is not brittle, dries hard in a short time, and does not require heating.

Since the beginning of the study in 1953, spherical material has been observed *in situ* only 36 times, once in a core of Miocene Sisquoc claystone and the remainder in rock chips of Cretaceous age from a weathered regolith and a silty claystone body both of which are now at great depth. The weathered surface contains a high concentration of limonite and hematite and badly leached quartz, now dull and deeply pitted. The largest populations of the particulate spheres are also encountered in this zone as contrasted to very rare single individuals found in the residues of sedimentary rocks which are believed to have almost solid. The thinnest walls have been observed only in the metallic highly magnetic group 2. The intergrowth group 3, like group 1, is characterized by thick walls. Group 1 spheres are black, nonmetallic, (Pl. 1, figs. 3–6), slightly magnetic to moderately magnetic, have thick walls, and frequently contain single or multiple solid, metallic, brassy-yellow, highly magnetic, malleable cores. All cores analyzed chemically contained large amounts of iron; a few contained small amounts of nickel.

Single and multiple nonmetallic spheres have been observed with cores (Pl. 1, fig. 1; Pl. 2, fig. 1). The shells of such spheres, after removal of the cores, were either weakly magnetic or nonmagnetic, indicating that the magnetic attraction exhibited by the metallic cores when in the spheres was sufficient to overcome any dampening effect by the outer-shell material. Observation, at high magnification in transmitted light, of finely powdered shell material indicates it is opaque. Analyses reveal silica and a small amount of iron and titanium.

In reflected light at 144 magnification the exteriors of the smooth nonmetallic shells of group 1 do not reflect light as a unit but as a surface of many iridescent specks probably caused by the presence of microscopic pits, each of which causes dispersion of a beam of



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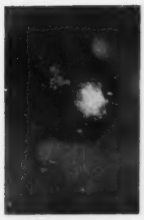

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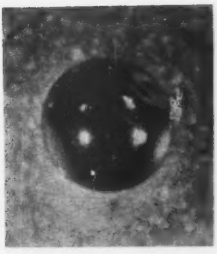

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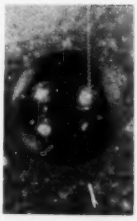

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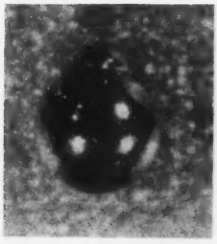

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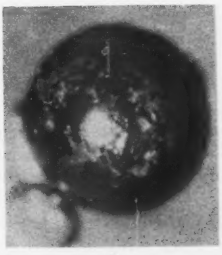
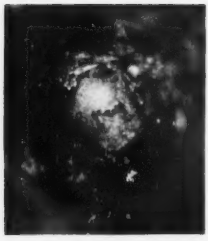


9

ANCIENT METEORITIC DUST

(1) Multiple composed of equal-sized group 1 spheres, nonmetallic shell of one partially removed to show metallic core now oxidized (X40)
(2) Group 2 sphere showing small subsidiary sphere on larger individual (X40)
(3, 4) Group 1 individuals, black, nonmetallic, smooth (X32)
(5) Group 1 individual, black, nonmetallic, smooth, with small subsidiary sphere (X32)
(6) Group 1 individual, black, nonmetallic faceted (X32)
(7–9) Group 3 individuals showing intergrown black and silver-gray exteriors. Figure 9 not photographed with diaphragm (X40)



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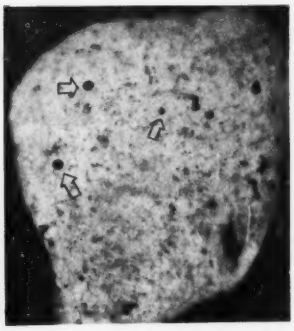

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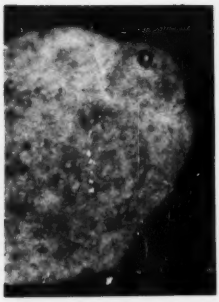

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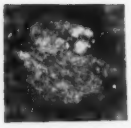

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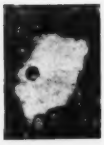

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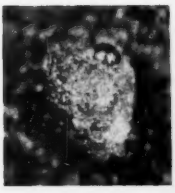

7

ANCIENT METEORITIC DUST

(1) Large single group 1 core-bearing sphere with nonmetallic shell partially removed (X36)
(2) Group 4 spheres (arrows) *in situ* in a chip of Upper Cretaceous clayey siltsone (X19)
(3, 6) Group 1 spheres *in situ* in a chip of Upper Cretaceous clayey siltstone (X19)
(4, 7) Group 3 spheres in hematitic cemented sandy siltstone from Upper Cretaceous regolith (X19)
(5) Group 3 sphere in hematitic cemented sandy siltstone from Upper Cretaceous regolith (X19)

white light into the spectrum. The exteriors of the faceted spheres in this group are composed of polygonal plates that give the appearance of crystal faces superimposed on a sphere (Pl. 1, fig. 6).

Group 2 is composed of hollow metallic, thin-walled spheres in the steel-gray to blue steel range. These are the most delicate spheres. Only a few tested contained nickel. The smooth individuals have highly polished exteriors, exhibit no detail at 144 magnification, and appear to be homogeneous. Others appear to be burnished, an effect produced by many groups of parallel fine striations. A few spheres in this group are clearly reticulate and closely resemble balls of twine.

Group 3 is the most populous and the most spectacular (Table 2). The spheres consist of intergrowths of steel-gray metallic and black nonmetallic material. Figures 7–9 of Plate 1 illustrate the details of the exteriors. Sectioning reveals that the walls contain many microscopic bubbles. The black plates at the surface are prismatic and appear to be perpendicular to and extend into the interior of the sphere. The steel-gray material is cryptocrystalline.

All the spheres sectioned were brittle, and it was impossible to maintain a plane surface during grinding. In group 3 spheres only the black plates maintained a plane surface; the metallic steel-gray material disintegrated into cryptocrystalline fragments and caused the unevenness of the grinding surfaces.

According to Handy and Davidson (1953) fly ash closely resembles meteoritic dust and is the noncombustible part of the powdered coal which escapes into the atmosphere during burning. It is composed of predominantly spherical, magnetic, gray, inorganic, glassy bodies composed of iron-alumina silicate. They estimate that approximately 7 million tons of this material enters the earth's atmosphere each year. The coarser fraction, which is electrostatically precipitated before escaping into the atmosphere, is used in well cementing.

Pokorny (1950, p. 79) concluded that microscopic glassy bodies and magnetic spheres observed in residues from cuttings resulting from drilling in Tertiary Flysch were contaminants from the smoke of coal-burning locomotives which had fallen into the drilling equipment while it was being hauled on freight cars.

Buddhue (1950, p. 12–22) indicates that most of the continents and ocean basins have yielded this material. Predominantly the reports have been concerned with the material falling from the atmosphere in historic time, but some have believed that in the past the earth was subjected to meteorite showers. Although Tarr (1932, p. 17–18) could find no evidence of prehistoric meteors he believed that under the proper conditions of burial in sediments a sealed environment would result, permitting the preservation of some of the meteoritic material. Buddhue (1950, p. 37) feels that ancient red clays and oozes now exposed in various parts of the world and ranging in age from Jurassic to Miocene probably contain these magnetic spherical meteoritic dust particles. He speculates that some of the spheres found in deep-sea cores were deposited from 10,000 to millions of years ago.

Group 1 of the California spheres is similar to material discussed by Buddhue (1950, p. 29), particularly his groups 4 and 5. These are the smooth, shiny to dull, nonmetallic spheres which in the California material usually contain malleable metallic cores. Buddhue (1950, p. 33) refers to metallic cores found in dull, black deep-sea spheres.

The material under study bears no resemblance to the fly ash described by Handy and Davidson (1953, p. 374), but it is similar to weld spatter and grinding dust. The non-

Table 2. Population of Spheres by Group

Group	Total number	Total range (mm)	Most populous range (mm)	Percentage in most populous range
1	49	0.119–0.951	0.119–0.951	100*
2	41	0.170–0.612	0.170–0.425	90.4
3	131	0.119–0.748	0.238–0.510	83.5
4	3	0.067–0.510	0.067–0.510	1.0

*22% occurred in the 0.340 mm size

magnetic fraction of the weld spatter, totaling 25 per cent of all weld debris, is composed of nonmetallic slag, is brown to black, transluscent to opaque, glassy, and is on the whole roughly similar to California group 4.

Three general groups make up the magnetic weld spatter spheres: a cindery scoriaceous group with no counterpart in the California classification; a smooth, highly polished, burnished, reticulate group similar to California group 2; and a third highly populous group containing many individuals that are exact duplicates of spheres in California group 3. Only one sphere in all the weld material studied was similar to California group 1. Grinding dust is composed entirely of magnetic, burnished to highly polished, highly magnetic spheres and is similar to California group 2.

Fly ash was rejected as a source material because of its dissimilarity; grinding dust was rejected as a source because of the rarity of the process at the well site. Consequently the study was narrowed to ancient and modern meteoritic dust and to weld spatter.

The discussion is limited to the spheres observed *in situ* in the Miocene and Cretaceous rocks. These are not contaminants, and their similarity to material attributed to extraterrestrial sources leaves little doubt as to their meteoritic nature. The presence or absence of nickel could not be regarded as an important factor since this element is used for many kinds of welding rod. Since some rod contains high percentages of titanium, isolated spheres rich in this metal are regarded as possible weld spatter.

The abundance of particulate material in association with spheres in rock chips in the Cretaceous regolith also favors meteoritic origin. More than 50 per cent of all the spheres were found in this zone. All the finds of high frequency have been associated with old subaerial surfaces on which deposition was sporadic or minor; in contrast, the shelf-type sediment yield has been on the order of 5 per cent of the total. The abundance of spheres on these old exposed surfaces is not mere coincidence, for deposition from the atmosphere onto a terrestrial surface would result in a higher concentration of meteoritic debris per rock volume than would deposition through the water column in a marine environment.

The only possibility of dilution on an exposed land surface would be if the meteoritic debris were entrapped during transport or deposition of sediments, but this possibility is far more remote than settling from the atmosphere into a marine environment characterized by heavy sedimentation. In this instance the spheres would not only be diluted by mixing with large volumes of settling terrestrial clastics but would also be dispersed by the turbulent and current action of the water column; the result would be the observed rarity in the shelf-type sediments.

The remainder of the evidence is more inferential. All four groups in the California classification are represented by spheres *in situ*. Although there is a general similarity of groups and frequency, only groups 3 and 4 are specifically similar by population and exterior to both types of material.

In the weld spatter the scoriaceous and the group 4 spheres make up a large proportion of the total, but none of the former material and only 1 per cent of the latter was found in the California material. Conversely group 1 was represented by only one individual after thousands of observations of weld spatter.

The similarity of both classes in groups 3 and 4 is not unusual when one considers that the mechanics of origin are similar. Buddhue (1950, p. 69), speaking of the origin of meteoritic dust, states:

> "Hot gases rushing past the melting surface would 'wipe off' the molten material in the form of filaments and irregular droplets. Surface tension would immediately break up filaments into droplets and cause the contraction of irregular droplets toward spherical form."

The study of weld spatter indicates that the transition from the molten stage to a solidified spherical individual is extremely rapid. Many of the bodies observed were perfect spheres. This requires complete solidification before the spatter came to rest. If not, at least one plane surface would be present as evidence of impact with the ground. Comparison of individuals of groups 3 and 4 of meteoritic dust with weld spatter indicates that solidification and concomitant development of the detailed exteriors of the meteoritic material were rapid.

It is postulated therefore that much of the California material is meteoritic in origin and that the earth has been subjected to the fall of meteorite showers since at least Late Cretaceous time and that this material can withstand weathering conditions severe enough to cause laterization. Identification of this material in well cuttings is hazardous because of composi-

tional and external similarity with weld spatter. Unless one observes the spheres *in situ* or in association with the proper environment, such as an ancient regolith, one must regard such particulate material with suspicion.

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